GENERAL CLIMATOLOGY

THIRD EDITION

GENERAL CLIMATOLOGY

HOWARD J. CRITCHFIELD
Professor of Geography
Western Washington State College

Prentice-Hall, Inc., Englewood Cliffs, New Jersey

Library of Congress Cataloging in Publication Data
CRITCHFIELD, HOWARD J
General climatology.

Bibliography: p.
1. Climatology. I. Title.
QC981.C73 1974 551.6 73-21719
ISBN 0-13-350264-3

© 1974, 1966, 1960 BY PRENTICE-HALL, INC.
ENGLEWOOD CLIFFS, NEW JERSEY

All rights reserved. No part of this book may be reproduced in any form or by any means without permission in writing from the publisher.

Printed in the United States of America

10 9 8 7 6 5 4 3 2 1

PRENTICE-HALL INTERNATIONAL, INC., LONDON
PRENTICE-HALL OF AUSTRALIA, PTY. LTD., SYDNEY
PRENTICE-HALL OF CANADA, LTD., TORONTO
PRENTICE-HALL OF INDIA PRIVATE LIMITED, NEW DELHI
PRENTICE-HALL OF JAPAN, INC., TOKYO

Contents

PREFACE, *xi*

1793985

PART ONE
The Physical Elements of Weather and Climate

one
CLIMATE AND THE ATMOSPHERE, *3*

The scope of climatology, 4
Nature of the atmosphere, 5
Origin of the atmosphere, 6
Composition of the atmosphere, 6
Vertical Divisions of the atmosphere, 8

two
HEAT AND TEMPERATURE, *12*

The radiation budget, 12
Variability of insolation, 15
Measurement of sunshine and insolation, 18
World distribution of insolation, 20
Air temperature and its measurement, 21
Thermometers, 23
Temperature records, 25
Horizontal temperature distribution, 27
Vertical distribution of temperature, 32

three
ATMOSPHERIC MOISTURE, *35*

Water vapor, 35
Humidity measurements, 37
Physical changes of state of water, 39
Processes of cooling to produce condensation and sublimation, 43
Clouds: Their formation and classification, 45
Cloud observations, 51
Fog, 53
Precipitation: Causes, forms, processes, and types, 56
Observations of precipitation, 60
Distribution of precipitation regionally, 61
Seasonal variation of rainfall, 64

four
MOTION IN THE ATMOSPHERE, *68*

Atmospheric pressure, 68
Pressure-height relationships, 71
Horizontal pressure distribution, 72
Pressure and wind, 76
Factors affecting wind direction and speed, 77
Winds aloft, 79
Diurnal variations of wind speed, 80
Wind observations, 80
Mapping wind data, 84
Local winds, 86
General atmospheric circulation, 88
Upper-level waves and jet streams, 90
Seasonal changes in the general circulation, 93
The monsoons, 96
Oceanic circulation, 97

five
WEATHER DISTURBANCES, *100*

Properties of air masses, 100
Air mass identification and analysis, 101
Stability and instability, 104
Air mass source regions and classification, 106
The extratropical cyclone, 109
The warm front, 111
The cold front, 112
Occluded fronts, 113
Mountain barriers and surface fronts, 115

Anticyclones, 116
Tropical storms, 117
Thunderstorms, 121
Tornadoes and waterspouts, 124
Special air mass and storm effects, 127
Storm and air mass observations, 130
The weather map, 132
Regional weather patterns, 134

PART TWO
Climates of the World

six
CLIMATIC CLASSIFICATION, *141*

Approaches to climatic classification, 141
Koeppen's classification, 145
The Thornthwaite classifications, 148
Other classifications, 152
Climatic regions of the world, 153

seven
CLIMATES DOMINATED BY EQUATORIAL AND TROPICAL AIR MASSES, *156*

The rainy tropics, 156
Monsoon tropics, 160
Wet-and-dry tropics, 164
Tropical arid and semiarid climates, 168

eight
CLIMATES DOMINATED BY TROPICAL AND POLAR AIR MASSES, *174*

Dry summer subtropics, 174
Humid subtropics, 178
Marine climate, 182
Mid-latitude arid and semiarid climates, 188
Humid continental warm summer climate, 192
Humid continental cool summer climate, 196

nine
CLIMATES DOMINATED BY POLAR AND ARCTIC AIR MASSES; HIGHLAND CLIMATES, *202*

Taiga climate, 202
Tundra climate, 206
Polar climate, 209
Highland climates, 212

PART THREE
Weather, Climate, and Life

ten
CLIMATE, VEGETATION, AND SOILS, *221*

Climatic factors in plant growth, 221
World distribution of vegetation, 225
Forests, 226
Grasslands, 228
Desert, 230
Tundra, 231
Vertical differentiation of vegetation, 233
Climate as a factor in soil formation, 235
Soil classification, 238
Spatial patterns of soils, 239
Soil erosion, 242

eleven
CLIMATE AND WATER RESOURCES, *246*

The hydrologic cycle, 247
The water budget at the earth's surface, 249
Evapotranspiration, 250
Soil moisture and ground water, 252
Runoff and floods, 254
Climatic causes of floods, 255
Runoff forecasting, 258
Snow surveying, 260

twelve
CLIMATE AND AGRICULTURE, *265*

Climatic factors in crop production, 266
Temperature and crops, 266
Frost, 268
Frost prevention, 271
The frost-free season, 274
Phenology, 275
The moisture factor, 280
Drought, 281
Combating drought—irrigation, 284
Crops and wind, 286
Climate and forestry, 288
Forest-fire weather, 292
Climatic factors in animal husbandry, 293
Selection and breeding for climatic adaptation, 297
Insects and diseases, 297
Plant and animal introduction, 300

thirteen
CLIMATE, TRANSPORTATION, AND INDUSTRY, *303*

Aviation, 303
Flying weather, 306
Turbulence, 308
Aircraft icing, 309
Flying in storms, 310
Water transport, 311
Railways and highways, 315
Public utilities, 318
Manufacturing, 321
Atmospheric pollution, 322
Construction, 326

fourteen
CLIMATE AND HOUSING, *330*

Climatic aspects of site, 330
Climate conditioning, 333
Building orientation, 333
Climate conditioning through design, 334
Air conditioning, 340
Heating, 341
Cooling, 345

fifteen
CLIMATE AND HUMAN COMFORT, *347*

Heat budget of the body, 348
Cooling power, 350
Clothing and climate, 353
Weather and health, 355
Sunshine and health, 357
Air pollution and health, 359
Climate and disease, 361
Acclimatization, 363
Perception of climate, 365

sixteen
CLIMATIC CHANGE, *368*

Paleoclimatology, 369
Geochronology of climates, 370
Climate during recorded history, 375
Recent climatic trends, 376
Climatic cycles, 377
Theories of climatic change, 378
Forecasting climate, 382

seventeen

MODIFICATION OF WEATHER AND CLIMATE, *384*

Modifying microclimates, 384
City climates, 385
Inadvertent modification of macroclimates, 389
Theoretical bases of planned modification, 392
Cloud modification, 397
Modification of storms, 401
Fog dispersal, 402
Implications of weather and climate modification, 404

APPENDICES

Appendix A Conversion equivalents, 409
Appendix B Abridged meteorological tables, 411

BIBLIOGRAPHY, *425*

INDEX, *429*

Preface

Travelling through space at more than 100,000 kilometers per hour in a nearly circular orbit around the sun, we have a vital need to understand the functioning of our planetary air-conditioning system. This book deals with the processes that involve heat, moisture, and motion in the earth's atmosphere. Its objective is to introduce the fundamentals of general climatology in a manner that will serve those who have a concern for the global environment, whether their primary interests lie in the natural sciences, the social sciences, or the emerging environmental sciences. Its scope encompasses both the nature of climate and the significance of climate to mankind.

This third edition retains the basic outline of the first and second editions and strives as before to incorporate recent developments in the sciences that deal with the atmosphere. A new chapter at the end summarizes advances in the growing study of inadvertent and experimental modification of weather and climate. There are also new sections on the water budget, the heat budget of the human body, air pollution and health, and perception of climate. The division of the broad field of climatology into three parts allows flexibility of both the emphasis and the topical sequence in a college course to fit various instructional patterns. Part One introduces the atmospheric exchanges of heat, moisture, and momentum that result in our weather and climate. It includes brief discussions of observational techniques, which are necessary for our understanding of atmospheric processes and regional differences in climate. Part Two treats the classification of climatic types and their geographical distribution, employing the fundamental pattern of world climates as a model for explanatory description. Part Three relates atmospheric processes to the biotic environment and human activities, concluding with chapters on past climatic changes and human modification of climates. Its arrangement facilitates treatment of separate sections in conjunction with topics in Parts One and Two if desired.

Both the text and figures of this edition emphasize the use of international units of measurement. If a justification is needed, it can be found anywhere in the realm of science and in the transactions of an increasing proportion of the world's people. The selections of supplementary readings at the ends of chapters and the bibliography are intended to encourage further investigation. They should guide the curious to sources of current information on topics of special interest.

I am greatly indebted to my teachers, who laid the foundation that made this book possible, and to my students, who have inspired many of the decisions on content and organization. I am especially grateful for the constructive criticism by teachers who have used previous editions in their classes. Any errors, inconsistencies, or omissions remain my responsibility, however.

HOWARD J. CRITCHFIELD

Bellingham, Washington

PART ONE

The Physical Elements of Weather and Climate

chapter one

Climate
and
the Atmosphere

Studies of the planet we call earth fall into four broad categories that embrace the solid *lithosphere,* the liquid *hydrosphere,* the gaseous *atmosphere,* and life forms of the *biosphere.* Although the study of weather and climate focuses on the envelope of gases, continuous interchanges among the "spheres" produce an integrated environment, and no component can be understood adequately without reference to the others.

The processes of exchange of energy and mass between the earth and the atmosphere over a long period of time result in conditions which we call *climate.* Climate is more than a statistical average; it is the aggregate of atmospheric conditions involving heat, moisture, and air movement. Extremes must always be considered in any climatic description in addition to means, trends, and probabilities. Climate is an important element of the physical environment of mankind, for, although man usually thinks of himself as a creature of the land, he actually lives at the bottom of a deep "ocean" of air that surrounds the earth. *Weather* is the day-to-day state of the atmosphere and pertains to short-term changes in conditions of heat, moisture, and air movement. Weather results fundamentally from processes that attempt to equalize differences in the distribution of net radiant energy received from the sun.

THE SCOPE OF CLIMATOLOGY

Climatology is the science that seeks to describe and explain the nature of climate, why it differs from place to place, and how it is related to other elements of the natural environment and to human activities. The term originated from the Greek words *klima,* referring to the supposed slope of the earth and approximating our concept of latitude, and *logos,* a discourse or study. Climatology is closely allied with *meteorology* (literally a "discourse on things above"), which treats day-to-day atmospheric conditions and their causes. Often defined as the physics of the atmosphere, meteorology uses the methods of the physicist to interpret and explain atmospheric processes; it is equated increasingly with atmospheric science. *Aerology* and *aeronomy* ordinarily denote studies of the upper atmosphere.

Climatology extends the findings of meteorology in space and in time to encompass the entire earth and periods of time as long as observations or scientific inference will permit. Because climatology involves the collection and interpretation of observed data, whether for developing generalizations or for testing theories, it necessitates instrumental and statistical techniques. It is more than statistical meteorology, however. In its investigations of spatial relationships it makes abundant use of the tools of geography, including maps. Its close affinity with geography arises from its multiple-factor approach to explanation of phenomena and from its concern with climate as an element of our natural environment.

The study of climate in the following chapters comprises three fundamental subdivisions: physical, regional, and applied. The basic question to be answered in physical climatology is: What causes the variations in heat exchange, moisture exchange, and air movement from time to time and from place to place? That is, why do we have different climates? As a first step toward the solution, we seek, through observation, facts relating to these variations. Several observable elements aid in the description of atmospheric conditions: radiation, duration of sunshine, temperature, humidity, evaporation, cloudiness and fog, precipitation, visibility, barometric pressure, and winds. The occurrence of these elements in a particular combination is the result of a number of processes involving the transfers of heat, moisture, and momentum which occur within the atmosphere and between the atmosphere and the earth's surface. Processes of weather and climate are influenced, in turn, by differences in latitude, altitude, land and water surfaces, mountain barriers, local topography, and such gross atmospheric features as prevailing winds, air masses, and semipermanent pressure centers. Part One discusses the roles of these factors and processes.

Regional climatology has as its goal the orderly arrangement and explanatory description of world climates. It includes the identification of significant climatic characteristics and the classification of climatic types, thus providing a link between the physical bases of climate and the investigation of problems involving climatic processes. Because it deals with spatial distributions regional climatology implicates the concept of scale. In keeping with the broad approach of this book, Part Two treats the large-scale macroclimatic regions of the world, but in a more detailed analysis mesoclimates and microclimates also require differentiation and explanation. Indeed, many of the problems whose study calls upon the principles and methodology of climatology are highly localized. As a result it is common to regard microclimatology as a group of techniques in applied climatology rather than as a branch of regional climatology. Where local climates are closely related to the surface conditions the term *topoclimatology* is sometimes employed to designate studies in an intermediate or mesoscale.

The third major division, applied climatology, explores the relationship of climatology to other sciences and considers its potential for improving human welfare, finally confronting the complexities of climatic change and modification of climates. It is in this phase of *general* climatology that the interdependence of sciences and the unity of human knowledge become increasingly evident. Not only has there been growing cooperation among the traditional sciences in the investigation of climatic problems, but new combinations also have begun to evolve. Bioclimatology, agroclimatology, medical climatology, building climatology, and urban climatology are examples. The proliferation of specialized studies clearly illustrates the wide scope and variety of climate-related phenomena.

NATURE OF THE ATMOSPHERE

The atmosphere is a deep blanket of gases and suspended liquids and solids that entirely envelops the earth. It is as much a part of the earth as the land and the oceans, but it differs from them in several respects. "Pure" air is colorless, odorless, tasteless, and cannot be felt except when in motion. It is mobile, elastic, compressible, expansible, and can transmit compression waves. It is transparent to many forms of radiation, yet it can absorb others. Although air is not nearly as dense as either land or water, it has weight and exerts pressure, but, because it is compressible, its density decreases rapidly with altitude. About half of the total mass of the atmosphere lies below 5,500 meters (18,000 feet), and nearly 99 percent lies within 30 kilometers (20 miles) of the earth's surface.

Fortunately, air offers resistance to objects which move through it so that the friction which results when meteors pass into the outer atmosphere creates enough heat to destroy most of them before they can reach the earth's surface.

Without the atmosphere life could not exist and there would be no clouds, winds, or storms—no weather. Besides being an essential for life and a medium for weather processes, the air acts as a great canopy that protects the earth from the full force of the sun by day and prevents the loss of too much heat at night.

ORIGIN OF THE ATMOSPHERE

The more plausible explanations of the creation of the atmosphere are based in part upon study of the atmospheres of other planets of the solar system. The elements which comprise the earth's atmosphere appear to be scarce elsewhere in the universe. Matter in the outer reaches of the atmosphere consists largely of the two lightest elements, hydrogen and helium, but these are rare in the air near the earth's surface. Presumably this condition developed through eons of geologic time along with slow changes that produced the present lithosphere and hydrosphere. In the early stages of formation of the planet from "cosmic gases" a weak gravitational field may have allowed light gases like hydrogen and helium to stray off into space. Even the gases of the present atmosphere probably are not a direct residue of the earliest form of the planet. Rather they are a secondary atmosphere which evolved from volcanic eruptions, hot springs, chemical breakdown of solid matter, and, later, contributions from vegetation. Evidence points to a stabilization of the atmosphere in something like its present form by the Cambrian period, about 580 million years ago.

Yet an interaction of land, water, air, and plant and animal life constantly uses and renews the atmosphere. For example, the weathering of rocks, burning of fuel, decay of plants, and breathing of animals all use oxygen and release carbon dioxide. Nitrogen follows a complex cycle through bacterial activity in the soil, animal tissue, organic processes in decay, and back into the air. Plants, animals, bacteria, and chemical interaction in soil and water all help to maintain an intricate balance among land, water, life, and the air.

COMPOSITION OF THE ATMOSPHERE

The air is a mixture of several gases, each of which acts independently of the others. The Greeks classed air as a single element, and that idea was held until the late eighteenth century when chemists began to isolate and identify separate gases. Simple experiments in chemistry and physics can be performed to demonstrate that air is neither a single gaseous element nor a compound of gases. Because of the continuous churning and diffusion of these gases in the lower atmosphere they are found in the same proportions at different times and places, so that the Greek concept of air is neither surprising nor entirely illogical.

Four gases—nitrogen, oxygen, argon, and carbon dioxide—account for more than 99 percent of dry air (Table 1.1). Nitrogen alone constitutes nearly four-fifths of the air by volume and oxygen one-fifth. The principal remaining stable gases are neon, helium, methane, krypton, hydrogen, nitrous oxide, and xenon. Various other less stable gases, including ozone and radon, also occur in air.

Of these gases oxygen is the most important for animal life. It combines readily with many other chemical elements and is necessary for combustion. Carbon dioxide is a product of combustion and is exhaled by animals and used by plants. It absorbs a part of long-wave radiation from the earth. Nitrogen does not combine readily with other elements, but it is a constituent of many organic compounds. One of its main effects in the atmosphere is to dilute oxygen and thus regulate combustion and other kinds of oxidation. Ozone is a powerful oxidizing agent but it occurs in such minute quantities and at such great altitudes that its function in this respect is extremely limited. Of greater significance is its capacity to absorb a part of the sun's ultraviolet radiation and thereby limit the quantity which reaches the earth's surface to the amount essential for life.

Up to this point we have considered the gases of dry air; however, of far more importance than any of them in weather and climate is water vapor which has characteristics that set it apart from the other gases of the air. The proportion of water vapor in the air varies from minute amounts to a maximum of about 4 percent under humid conditions near the earth's surface. This variability is one of the main concerns of the meteorologist. In Chapter 3 it will be discussed in detail along with another unique feature of water vapor, namely, its ability to change from

TABLE 1.1

Principal Gases Comprising Dry Air

	By Volume	*By Weight*
Nitrogen (N_2)	78.088%	75.527%
Oxygen (O_2)	20.949	23.143
Argon (A)	0.93	1.282
Carbon dioxide (CO_2)	0.03	0.0456
	99.997	99.9976

a gas to a liquid or solid while a part of the atmosphere. The three forms of water in the atmosphere have restrictive effects upon solar radiation; they are able to absorb, reflect, and scatter certain wave lengths of the solar spectrum.

Up to altitudes of 50 or 60 mi (80–100 km) the composition of the atmosphere remains fairly constant. The proportion of ozone increases to a maximum at about 25 km; carbon dioxide, water vapor, and dust decrease. Besides the water droplets and ice crystals which occur as fog, haze, or clouds in the atmosphere, there are large quantities of suspended dust particles. The amount of dust varies greatly over the earth, but even over the oceans, hundreds of particles per cubic centimeter of air have been counted. Most of these particles are invisible to the naked eye and some are submicroscopic in size. Dust absorbs a part of the incoming solar radiation and also is an agent of reflection and scattering. It is one of the factors in the intensity and duration of dawn and twilight. The blue color of the sky and the red of sunsets are due to selective scattering of the visible solar spectrum by gas molecules and dust. Condensation of water vapor begins on nuclei of fine dust particles. One might suppose that the atmospheric dust would eventually be washed down to earth by rains, but it is constantly being replenished. The sources are many and varied: dry soil particles, smoke, salts from ocean spray, bacteria, seeds and spores, volcanic ash, and meteoric dust. The number of dust particles is great in industrial centers and relatively greater in dry regions than in humid areas. The concentration of dust generally decreases with altitude although the smallest particles reach heights of several kilometers and meteoric dust is introduced at the outer limits of the atmosphere.

VERTICAL DIVISIONS OF THE ATMOSPHERE

The vertical extent of the atmosphere is difficult to ascertain, for there is no sharp boundary between the air and extraterrestrial space. Atmospheric phenomena associated with the earth's magnetic and gravitational fields extend outward for several thousand kilometers to a vague zone of nebulous gases and radiation particles that become rarer and rarer until at last terrestrial characteristics of the atmosphere cease.

On the basis of vertical temperature differences the atmosphere may be divided into four major layers, or "shells"—the *troposphere*, the *stratosphere*, the *mesosphere*, and the *thermosphere* (Figure 1.1). The troposphere is the lower portion of the atmosphere, extending up to about 8 km at the poles and 16 km at the equator. It contains three-fourths of the atmospheric mass and is the realm of clouds, storms, and

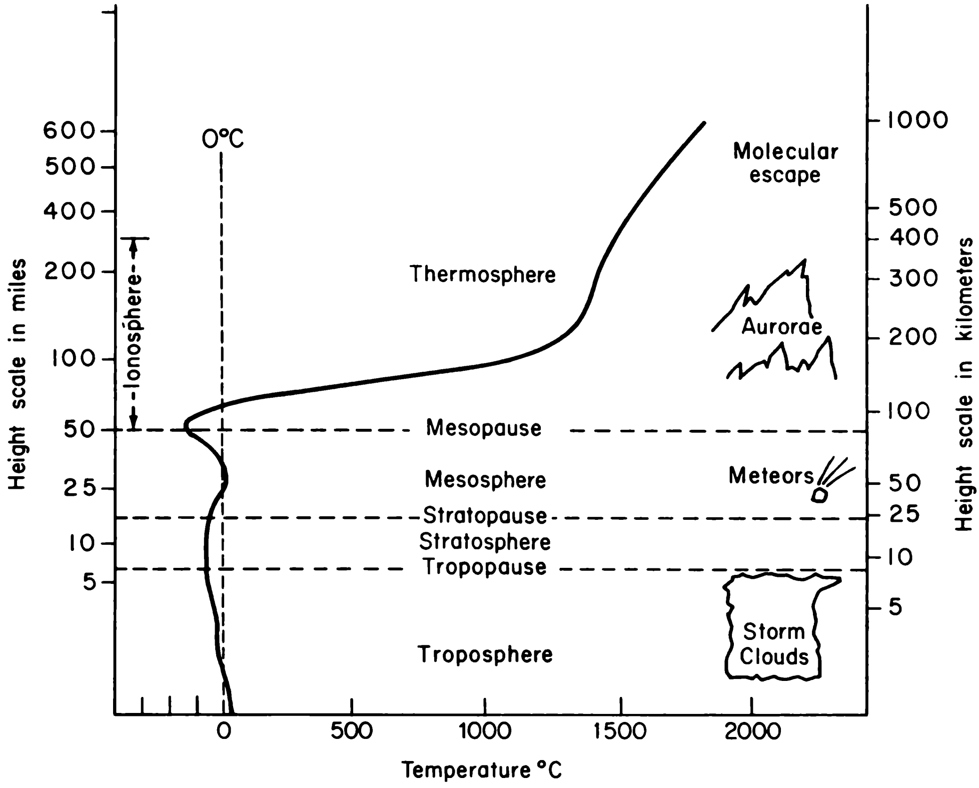


Figure 1.1
Vertical temperature divisions of the atmosphere and related phenomena.

convective motion. Thermal convection, being better developed in the tropics, is responsible for the greater vertical extent of the troposphere near the equator. It also explains the somewhat greater height of the troposphere in summer than in winter at a given latitude. The outstanding characteristic of the troposphere is the fairly uniform decrease in temperature with increase in altitude until minimums of −50° or −60°C are reached. The zone marking the end of this temperature decrease is called the *tropopause*. Above the tropopause lies the stratosphere, which exhibits little or no vertical change in temperatures that approximate those of the arctic winter at the surface. The stratosphere is thickest over the poles and sometimes does not exist over the equator. At its inner margin sulphate particles are concentrated. Near its outer limit, the *stratopause*, are the greatest concentrations of ozone, and rare nacreous (mother-of-pearl) clouds appear occasionally. Beyond the stratopause, which has a mean altitude of about 25 km in the middle latitudes, is the mesosphere. Temperatures in the mesosphere increase and then decrease to a minimum near the *mesopause*, at 80 km. Most meteors burn and disintegrate in this layer. In the thermosphere the temperature again rises to values approaching 1700°C at an undefined upper limit.

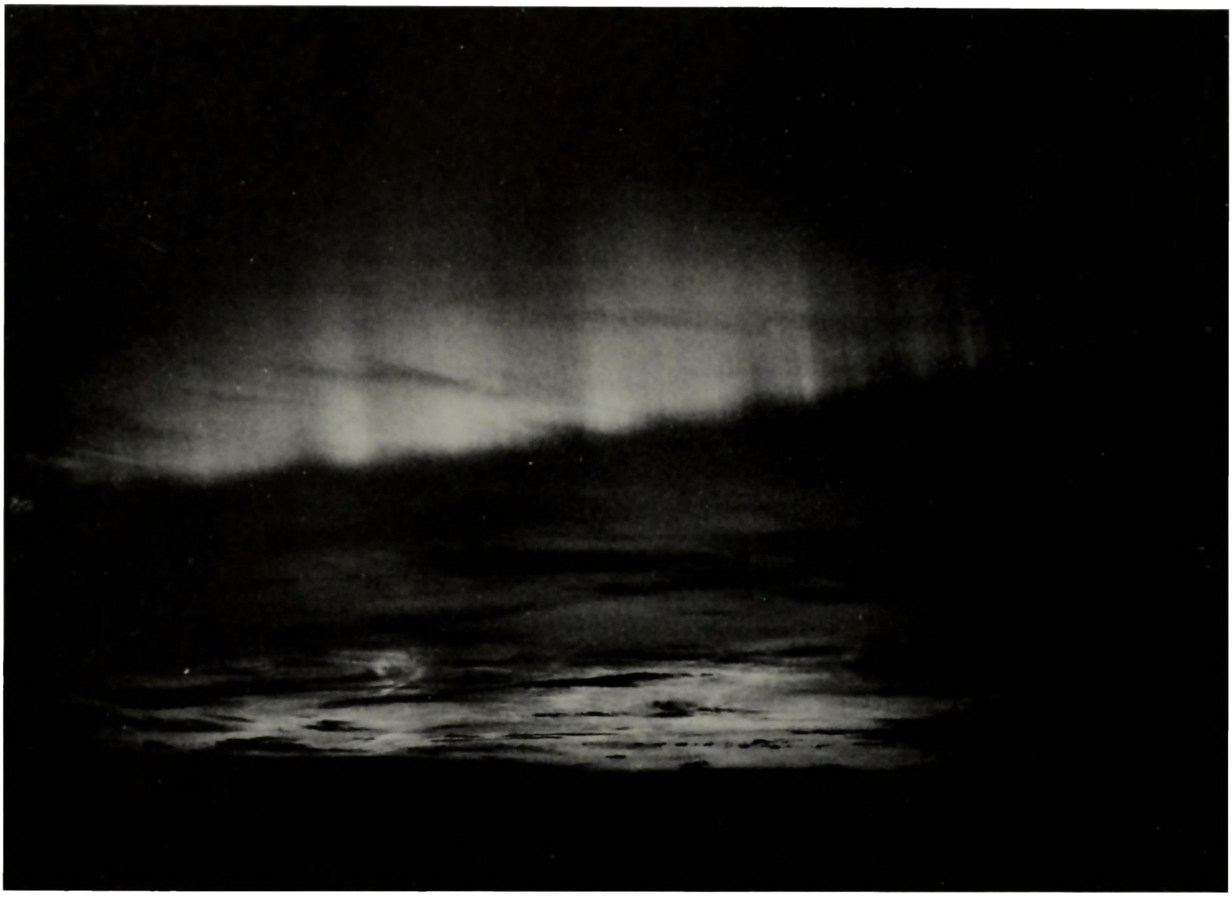


Figure 1.2
Aurora borealis and noctilucent clouds at Grande Prairie, Alberta, on the night of 18 July 1965. (Courtesy Benson Fogle.)

Such temperatures are not strictly comparable with those observed by thermometers at the earth's surface, however. Although the gas molecules exhibit high kinetic energy, and therefore have high temperatures, they are too sparse to transfer significant quantities of energy to an ordinary thermometer. The exposed hand of an astronaut would not feel hot in the thermosphere.

Coinciding with the lower portion of the thermosphere is the *ionosphere,* an atmospheric layer at 100 to 400 km delimited on the basis of ionized particles and their effects on the propagation of radio waves. Little was known of the ionosphere until it was found that radio waves are reflected by ionized layers at great heights. The aurora borealis (Figure 1.2) and its Southern Hemisphere counterpart, aurora australis, apparently result from excitation of the ionosphere by high-energy radiation particles from the sun. Their maximum occurrence is near the magnetic poles, toward which the particles are deflected by the earth's magnetic field. Auroras have been observed to increase with an increase in sunspot activity. There is no proven relationship between auroral displays and weather in the troposphere. Another phenomenon of the ionosphere, *air glow,* originates with self-luminescent gases. It exhibits no significant correlation with sunspot or magnetic activity.

In the chapters that follow we shall be concerned almost entirely with the lower part of the troposphere, for it is the sphere of weather, but the unity of the atmosphere must be kept in mind. Variations of density, composition, or temperature in upper atmospheric layers may logically be expected to produce changes in lower layers. Much of what is not now fully understood about our weather and climate may be explained ultimately with reference to the upper atmosphere.

Additional Readings

COURT, ARNOLD, "Climatology: Complex, Dynamic, and Synoptic," *Ann. Assoc. Am. Geogrs.,* 47, 2 (1957), 125–36.

FRISINGER, H. HOWARD, "Aristotle and His 'Meteorologica,'" *Bull. Am. Met. Soc.,* 53, 7 (1972), 634–38.

GOODY, RICHARD M., AND JAMES C. G. WALKER, *Atmospheres.* Englewood Cliffs, N.J.: Prentice-Hall, Inc., 1972.

HARE, F. KENNETH, "The Stratosphere," *Geog. Rev.,* 52, 4 (1962), 525–47.

HAURWITZ, B., "Atmospheric Tides," *Science,* 144, 3625 (1964), 1415–22.

JASTROW, ROBERT, "Artificial Satellites and the Earth's Atmosphere," *Scient. Am.,* 201, 2 (1959), 37–43.

LANDSBERG, H. E., "Origin of the Atmosphere," *Scient. Am.,* 189, 2 (1953), 82–86.

WALKER, J. C. G., AND N. W. SPENCER, "Temperatures of the Earth's Upper Atmosphere," *Science,* 162, 3861 (1968), 1437–42.

WEBB, WILLIS L., *Structure of the Stratosphere and Mesosphere.* New York: Academic Press Inc., 1966.

chapter two

Heat and Temperature

The sun is the primary source of energy for the processes of change at the earth's surface and in the atmosphere. The amount of energy that the earth receives from the stars and other celestial bodies is negligible. Radiant energy from the sun that strikes the earth is called *insolation*—a contraction of "incoming solar radiation." Insolation consists of rays of radiant energy made up of various wave lengths. The longer waves, infrared, are largely absorbed in the troposphere. At the other end of the solar spectrum the shorter ultraviolet waves heat the upper atmosphere and are capable of producing certain photochemical effects. Between these bands of invisible radiation is the visible portion of the spectrum that we know as sunlight and that is most effective in heating the earth. When the components of the solar spectrum strike the earth they are partially absorbed and converted from short-wave to long-wave radiant energy, that is, to heat. Energy gained in this way provides the "fuel" for the processes of weather and climate, and it is transferred both vertically and horizontally, creating a variety of temperature conditions. Eventually it is returned to space by the process of radiation.

THE RADIATION BUDGET

Only a small part of insolation is absorbed directly by the air through which it passes. The processes of radiation, convection, and conduction bring about heating and cooling of the air by transferring heat between the earth and the air and between different levels in the atmosphere. In addition, the processes of evaporation and condensation effect exchanges of heat between various surfaces and the overlying air.

Radiation is the process of energy transmission by electromagnetic waves and is the means by which solar energy reaches the earth (Figure 2.1). Of the total solar radiation incoming at the outer limits of the atmosphere about 31 percent is reflected by clouds or scattered back to space by clouds and dust and is not used to heat the air; 4 percent is reflected to space from the earth's surface. The ratio of the amount of radiation reflected to the amount received on a surface is termed the *albedo*. The earth's mean albedo is about 35 percent, although it is greater in polar regions than at the equator, varying with the angle of incidence as well as the color of surfaces. The albedo of clouds varies widely with their thickness and composition. On the average clouds cover about half of the globe, giving them a major influence on the radiation budget. Table 2.1 lists the albedos of several kinds of surfaces.

About 18 percent of insolation is absorbed by gases (principally water vapor), clouds, and suspended solids in the atmosphere; the earth absorbs 23 percent directly at its surface and 24 percent indirectly after

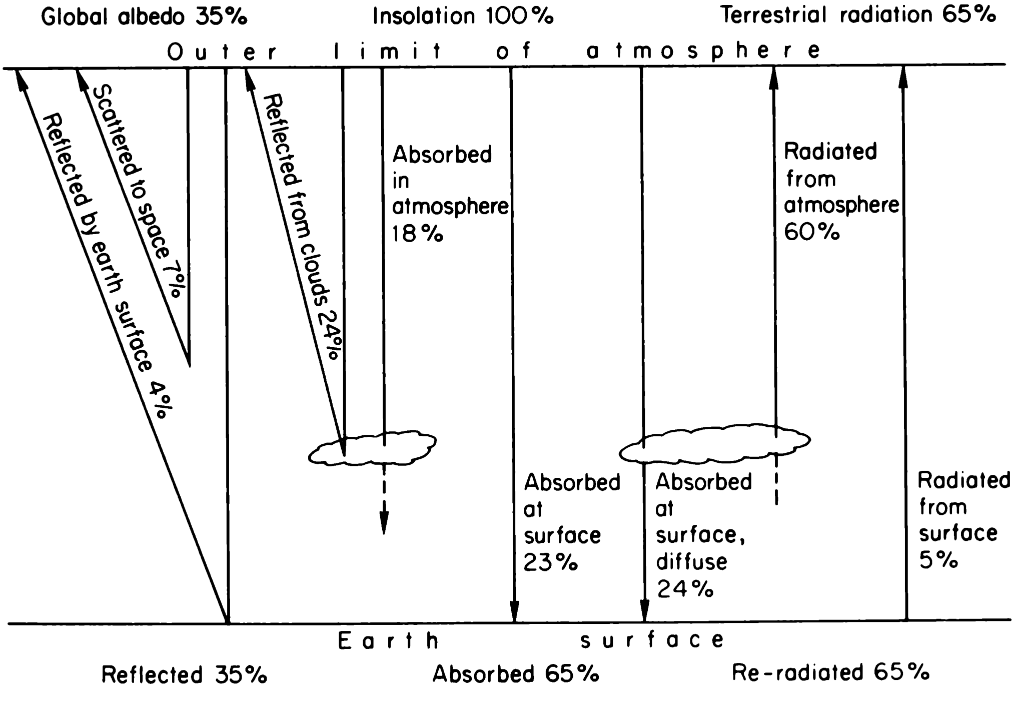

Global albedo 35%
Insolation 100%
Terrestrial radiation 65%
Outer limit of atmosphere
Reflected by earth surface 4%
Scattered to space 7%
Reflected from clouds 24%
Absorbed in atmosphere 18%
Radiated from atmosphere 60%
Absorbed at surface 23%
Absorbed at surface, diffuse 24%
Radiated from surface 5%
Earth surface
Reflected 35%
Absorbed 65%
Re-radiated 65%

Figure 2.1
Radiation budget of the earth and its atmosphere.

TABLE 2.1
Estimated Mean Albedos of Various Surfaces

Surface	*Albedo in Percent*
Tropical forest	21
Deciduous forest	18
Coniferous forest	13
Savanna	15
Desert	28
Oceans (60–70° lat.)	7–23
Inland waters	2–78
Snow	40–90
Wet sand	30–35
Bare rock	12–18
Grain crops	10–25
Green grass	8–27

diffuse scattering by clouds and the atmosphere. Thus, approximately two-thirds of the total insolation is effective in heating the earth and its atmosphere. As a result of absorption of insolation the earth itself becomes a radiating body that emits energy primarily in the long wave lengths. Although the components of the atmosphere collectively are able to absorb only about a fifth of the incoming short-wave radiation, they can capture a large part of the outgoing long-wave radiation. This ability of the atmosphere to admit most of the insolation and retard losses by reradiation from the earth is known as the *greenhouse effect.* Besides the long-wave (heat) radiation from the earth's surface there is also radiation from cloud layers and from the atmosphere out into space. The total amount of energy coming into the planet earth over a considerable period of time is equalled by total outward losses. If this were not so the earth would soon become either very hot or very cold. Actually there is a deficit of heat at high latitudes and a surplus in low latitudes owing to radiation gains and losses. The character of different surfaces also produces regional variations in the heat budget. Horizontal transport by winds and by ocean currents prevents the progressive accumulation of heat in equatorial regions or in areas having high absorptive qualities.

Conduction is the heat transfer process which occurs when two bodies of unequal temperature are in contact. Heat passes to the colder body as long as the temperature difference exists. Thus, when the land is absorbing radiation and is warmed above the temperature of the air, conduction transfers a part of the heat to the lower layer of air. When land is cooler than the adjacent air the heat transfer is reversed and the air is cooled. The latter phenomenon is common at night and in winter in the middle and high latitudes.

Air itself is a poor conductor of heat. If conduction alone provided transfer of heat upward from the earth's surface, the air would be very hot along the ground on a summer day and quite cool a few meters up. Within the air heat conduction is insignificant compared with *convection,* which accomplishes transfer of heat through movement of the air itself. Air heated by contact with the warm earth surface expands and becomes less dense. The lighter air thus created is replaced by cooler, heavier air from above and a convectional circulation is established with horizontal as well as vertical components. The same principle is involved in the circulation of warm air in a room. Liquids, being mobile like gases, also develop convectional systems. The oceans exhibit convection on a grand scale in their pattern of currents. Although much of the convective activity is due to heating at the earth's surface, it is important to remember that the principle of convection will operate also through cooling at higher levels. An example of this occurs when cold air temporarily pushes in above warmer air, producing a sudden overturning of air as the denser, cooler air above replaces the lighter, warmer air below. Winter thunderstorms sometimes result from this kind of convection.

Although part of the net radiation gained by the earth's surface is transferred upward by long-wave radiation and as sensible heat by the processes of conduction and convection, much of it reaches the atmosphere as latent heat in water evaporated from land and water surfaces, and especially from oceans. Moist air masses may move great distances before condensation of their water vapor releases huge quantities of latent heat. Thus we see that the processes of heat and moisture exchange are closely interconnected.

Taken together, the processes of heat transfer maintain a complex exchange of heat between the earth's surface and the atmosphere. The heat budget of a particular place is not often in balance. In middle or low latitudes a true balance may exist briefly near the beginning and end of each daylight period. Ultimately, however, the total amount of energy coming into the earth is equalled by total outward losses to space. Satellite measurements have confirmed a fairly stable global heat budget over relatively long periods. The potential effects of an imbalance upon climate will be treated in Chapter 16. Although the emphasis here has been on the entire globe, the heat budget concept applies equally to geographic regions, buildings, plants, animals, and even the human body.

VARIABILITY OF INSOLATION

The amount of insolation received on any date at a place on the earth is governed by:

1. the *solar constant,* which depends on:
    (*a*) energy output of the sun
    (*b*) distance from the earth to the sun
2. transparency of the atmosphere
3. duration of the daily sunlight period
4. angle at which the sun's noon rays strike the earth

The first of these influences on insolation is fundamental. Measurements and calculations of the amount of energy coming into the outer limits of the atmosphere have resulted in an average figure of about 2 langleys (i.e., 2 gram-calories per square centimeter) per minute.

Although referred to as the *solar constant,* it actually varies slightly. The variation has not been found to have a great effect upon daily weather, but it may be related to certain climatic fluctuations. Variations in specific components of the solar spectrum, especially ultraviolet radiation, are probably of greater significance in affecting atmospheric conditions.

The distance between the earth and sun varies between 152 million km at *aphelion* (July 4) and 147 million km at *perihelion* (January 3) (Figure 2.2). Neither of these is a great variation from the average distance of 149.5 million km (93 million mi), and although the amount of radiation reaching the outer atmosphere is about 7 percent greater at perihelion than at aphelion, other factors influencing insolation and temperature largely override the difference.

Transparency of the atmosphere has a more important bearing

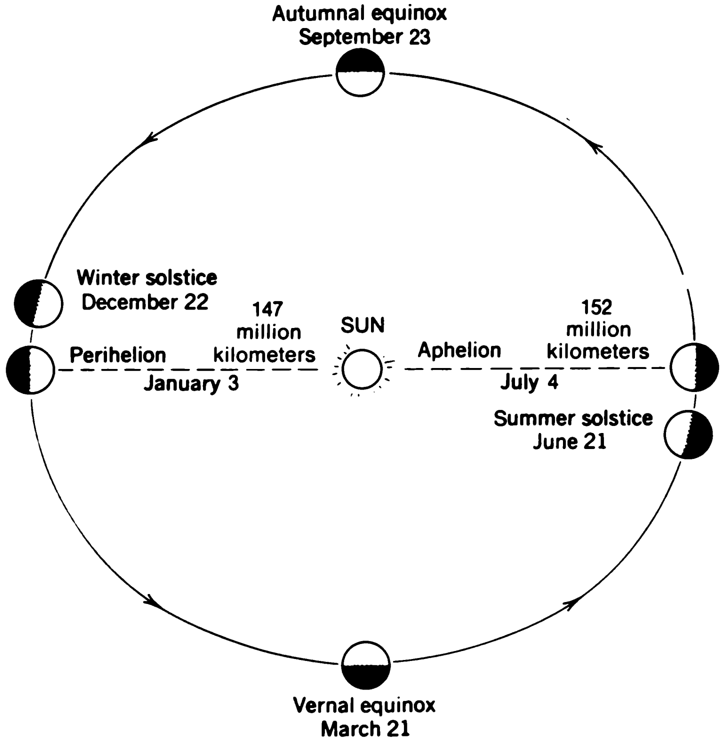


Figure 2.2
The earth's orbit around the sun.

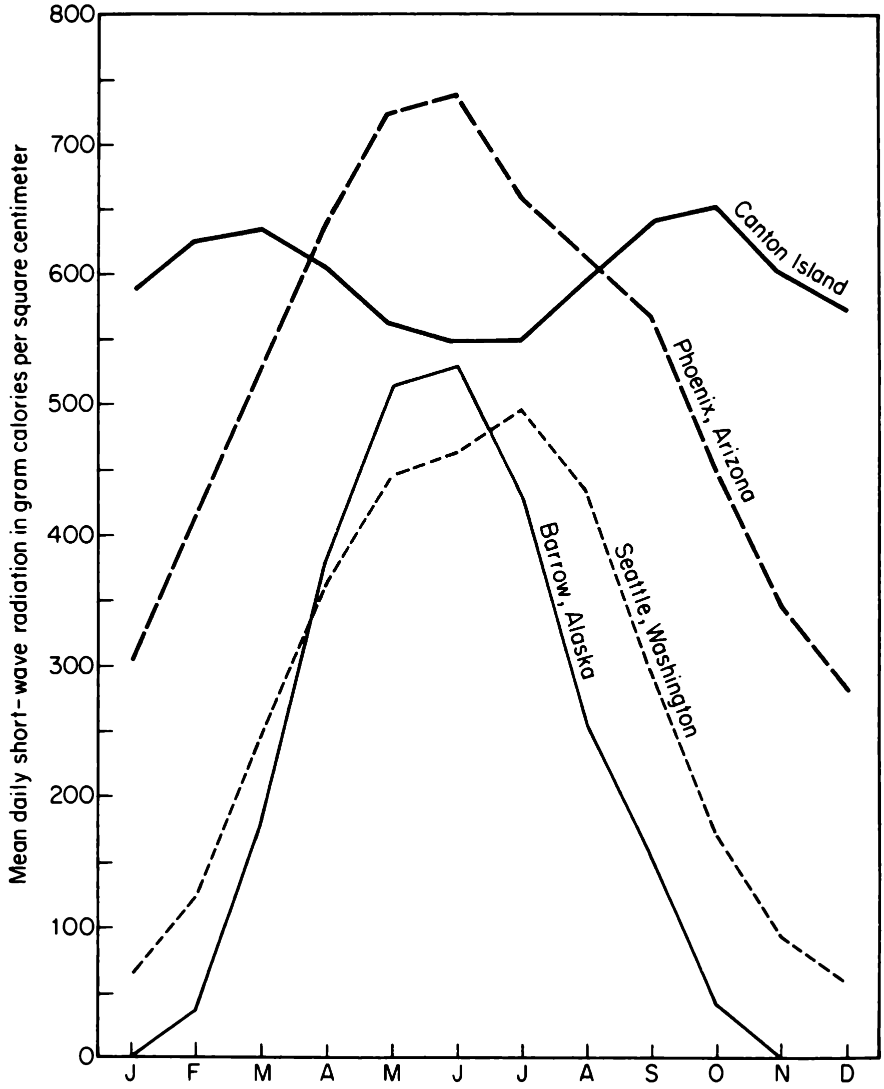


Figure 2.3
Annual march of mean daily short-wave solar radiation at selected stations.

upon the amount of insolation which reaches the earth's surface. The effect of dust, clouds, water vapor, and certain gases in reflection, scattering, and absorption was discussed previously. It follows that areas having dense cloudiness or polluted air will receive less direct insolation. Transparency is also a function of latitude, for at middle and high latitudes the sun's rays must pass through the reflecting-scattering atmosphere at a lower angle than in tropical latitudes. This effect varies with the seasons, being greatest in winter when the sun's rays are lowest on the horizon (Figure 2.3).

The duration of daylight also varies with latitude and the seasons,

TABLE 2.2
Longest Possible Duration of Insolation

Latitude	0°	17°	41°	49°	63°	66½°	67°21′	90°
Daylight	12 hr	13 hr	15 hr	16 hr	20 hr	24 hr	1 mo	6 mo

and the longer the period of sunlight the greater will be the total possible insolation. At the equator day and night are always equal. In the polar regions the daylight period reaches a maximum of 24 hours in summer and a minimum of zero hours in winter. At the respective summer solstice, under clear skies, a polar area may receive more radiation per 24-hour day than other latitudes, although the net radiation used for heating is reduced by the albedo of ice and snow surfaces.

The effect of the varying angle at which the sun's rays strike the earth can be seen in the daily march of the sun across the sky. At solar noon the intensity of insolation is greatest, but in the morning and evening hours, when the sun is at a low angle, the amount of insolation is small (Figure 2.4). The same principle has a broader application with respect to latitude and the seasons. In winter and at high latitudes even the noon sun's angle is low; in summer and at low latitudes it is more nearly vertical. The oblique rays of the low-angle sun are spread over a greater surface than are vertical rays and, as a consequence, produce less heating per unit area.

The angle at which solar radiation strikes the earth's surface also depends upon the surface configuration of the land. In the Northern Hemisphere southern slopes receive more direct rays, whereas northern slopes may be entirely in the shade. The possible hours of sunshine in a deep valley may be greatly reduced by surrounding hills. Certain valleys in Switzerland, for example, receive direct sunlight in winter for only two or three hours around noon.

MEASUREMENT OF SUNSHINE AND INSOLATION

The effects of sunshine as distinct from heat and temperature are important in maintenance of life, and they play a large part in human activities. Everyone is familiar with the claims of many chambers of commerce that theirs is a climate with a great number of sunshine hours.

There are several types of instruments which measure the duration of sunshine (see Figure 2.5). The Campbell-Stokes recorder focuses the sun's rays through a glass globe which, acting as a burning glass, chars a record of sunshine on a graduated strip of cardboard. Modified cameras have also been devised which use photographic paper to record the sun's light rays. Another class of sunshine recorder depends upon the fact that

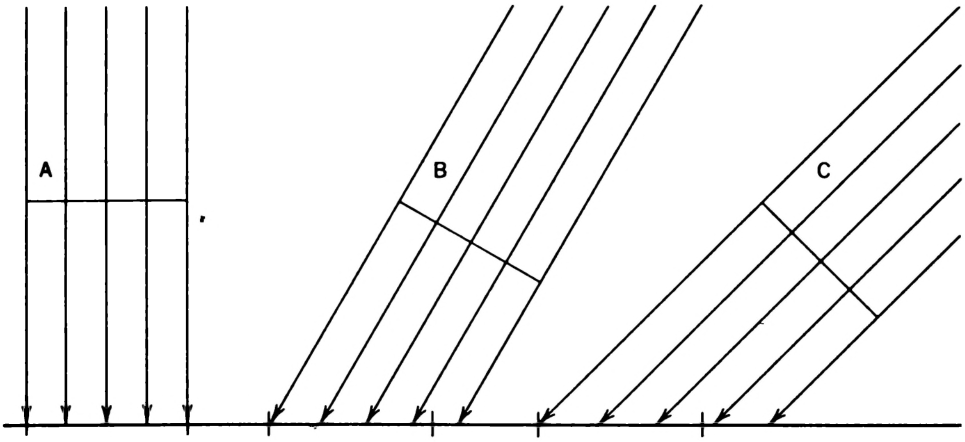
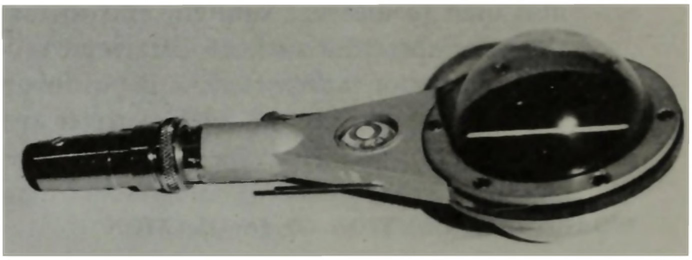
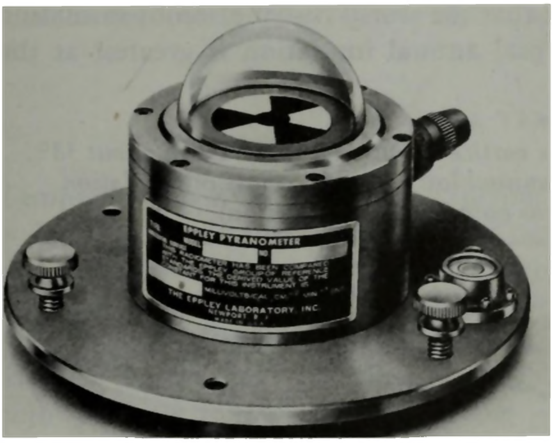


Figure 2.4
Relation of sun angle to effectiveness of insolation. A, *sun directly overhead, highly effective;* B *and* C, *smaller angles of incidence, less effective insolation per unit area.*

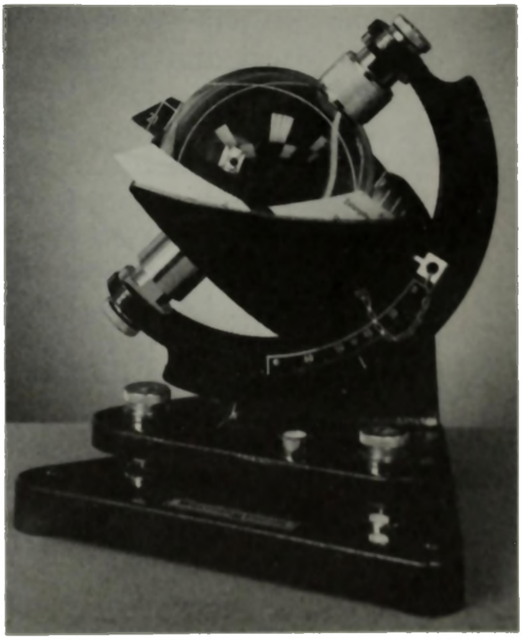


A



B

Figure 2.5
Instruments for measurement of radiation and sunshine. A, Campbell-Stokes sunshine recorder; B, Eppley pyranometer; C, net radiometer. (*A and C courtesy Weather Measure Corporation; B courtesy Eppley Laboratory, Inc.*)



C

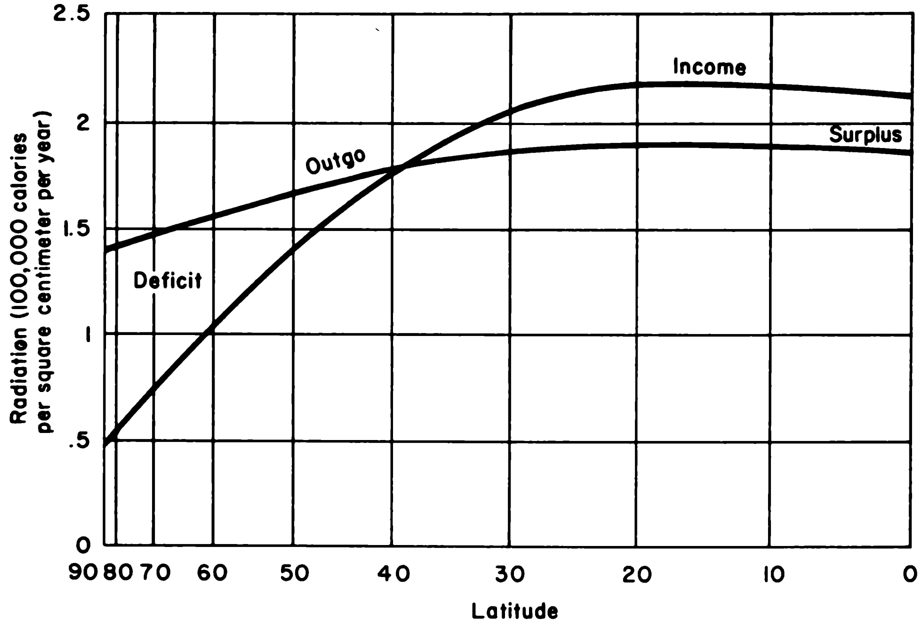
a black surface absorbs radiation at a faster rate than a white or translucent one. The resulting heat differential is transferred mechanically or electrically to a recorder.

Instruments for the determination of insolation, called *pyranometers,* measure the total amount of short-wave radiation absorbed on a horizontal surface. A common type operates on a thermoelectric principle. The incidence of radiation upon black and white surfaces of equal area produces differential heating that generates an electromotive force proportional to the intensity of radiation. The instrument is coupled electrically to indicating and recording devices. *Pyrheliometers* are essentially specialized pyranometers, shielded to intercept only direct solar radiation and oriented so that receiving surfaces are perpendicular to the solar beam. Photoelectric cells and photospectroscopic apparatus are also used to measure sunlight and to investigate special components of the solar spectrum such as ultraviolet radiation. By means of net radiation meters, or *radiometers,* it is possible to measure the radiant transfer of heat upward from the earth's surface as well as that directed downward.

WORLD DISTRIBUTION OF INSOLATION

From the foregoing it is evident that the world distribution of insolation is closely related to latitude. Total annual insolation is greatest at the

Figure 2.6
Mean latitudinal distribution of the earth's radiation budget. From about 38° N and S to the poles there is a net annual loss of heat by outgoing radiation. Between 38° N and S income exceeds outgo. (After Houghton)



equator and decreases toward the poles. At the equator the annual receipt of insolation is about four times that received at either of the poles (see Fig 2.6). As the rays of the sun shift seasonally from one hemisphere to the other, the zone of maximum possible daily insolation moves with them. In tropical latitudes the amount of insolation is constantly large, and there is little variation with the seasons. But in its annual journey the sun passes over all places between the Tropic of Cancer and the Tropic of Capricorn twice, causing two maximums of insolation. In the latitudes between 23½° and 66½°, maximum and minimum periods of insolation occur at the summer and winter solstices respectively. Beyond the Arctic and Antarctic Circles the maximum coincides with the summer solstices, but there is a period during which insolation is lacking. The length of the period increases toward the poles, where it is of six months' duration.

Observations of actual insolation at the earth's surface show a distribution that departs slightly from a simple latitudinal pattern. Maximum annual values of insolation occur at about 20° latitude, where the drier air permits a greater proportion of the radiant energy to penetrate to surface levels. Regions with considerable cloudiness receive less insolation at the surface than do areas with predominantly clear weather. In general, high plateaus and mountains are favored by more effective insolation because of the relatively clearer air at high altitudes (Figure 2.7).

AIR TEMPERATURE AND ITS MEASUREMENT

Temperature is a relative term implying a degree of molecular activity, or heat, of a substance. If the heat of one body flows to another we say that the former has the higher temperature. To measure the temperature of a body an arbitrary scale of reference is employed. The two most common scales used in the measurement of air temperature are the Celsius (or Centigrade) and Fahrenheit scales. The Celsius scale's boiling point of water is 100°, and its melting point of ice is 0°. It is the scale accepted internationally for reporting temperature data and for meteorological analyses. The Fahrenheit scale fixes the boiling point of water at 212° and the melting point of ice at 32°. The two scales indicate the same temperature at −40°. Celsius temperatures may be converted to their Fahrenheit equivalents by the following formula:

$$F = 32 + 9/5C$$

Similarly:

$$C = 5/9(F - 32)$$

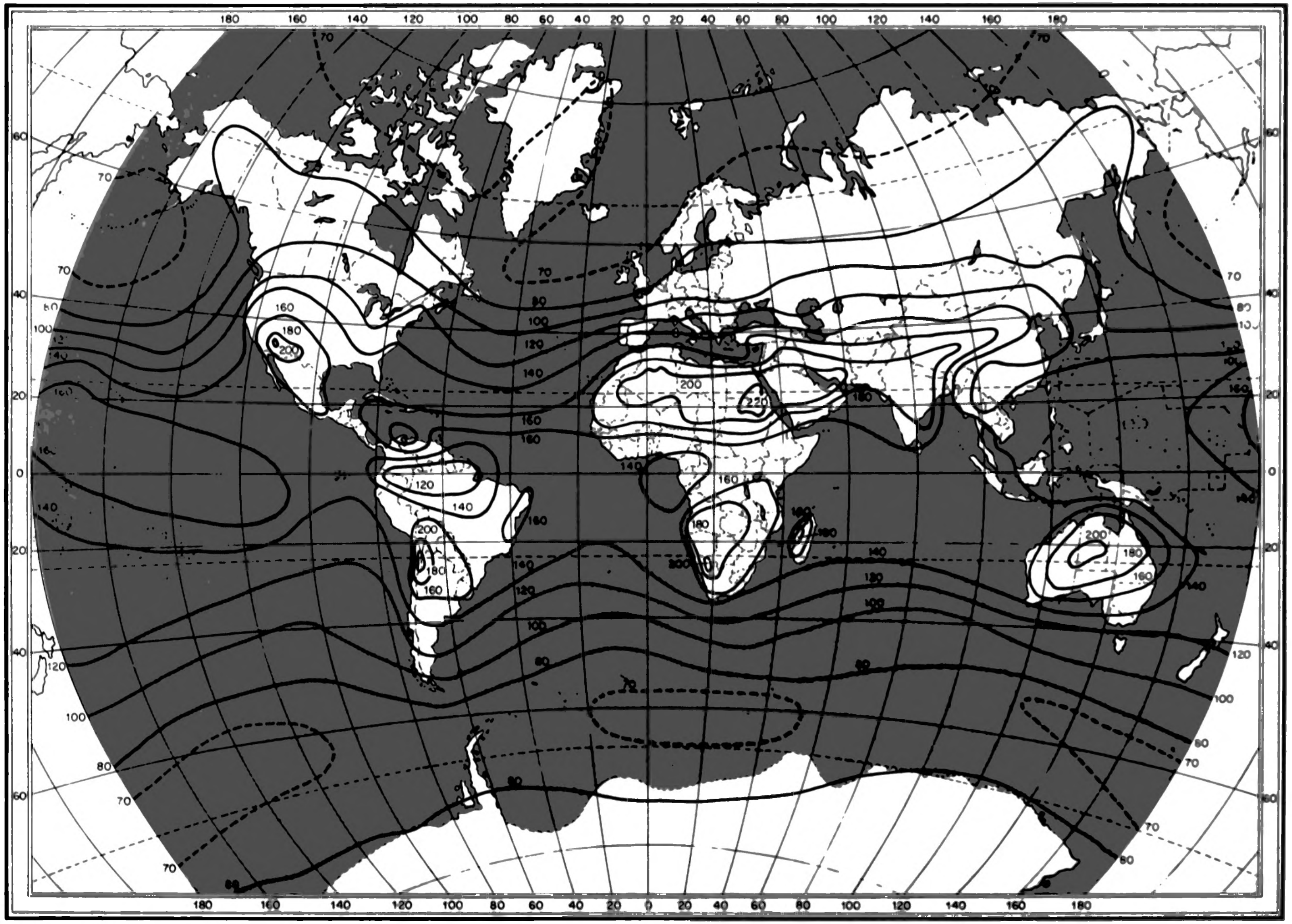


Figure 2.7
Mean annual total direct and diffuse radiation at the earth's surface. Values are in kilogram-calories per square centimeter per year on a horizontal surface.
(After Budyko and Landsberg; modified Van der Grinten projection)

Still another scale, the Kelvin, or Absolute, is based on *absolute zero,* the temperature at which a gas theoretically would cease to exert pressure. It is commonly used to indicate temperatures in the upper atmosphere and in calculations involving energy exchange. The value of each degree on the Kelvin scale equals that of a Celsius degree, but Kelvin (or absolute) zero is at −273.16°C. For ordinary purposes we can convert Celsius to Kelvin simply by adding 273.

THERMOMETERS

The most common type of thermometer consists of a sealed glass tube with a uniform bore and a bulb at one end filled with either mercury or alcohol. Changes in temperature produce expansion or contraction of the contents of the tube, which is graduated to facilitate reading of the indicated temperature values. The accuracy of temperature measurements depends on the care with which the thermometer is constructed and calibrated as well as on its exposure to the air. For official measurements of surface air temperature the thermometer is mounted in a louvered instrument shelter.

Another type of thermometer indicates temperature differences as they affect the shape of a bimetallic element. Two strips of different metals are welded together to form a bimetallic bar. An increase in temperature causes the bar to expand, but the two metals expand at different rates, causing the unit to bend. The curvature is translated into temperature values on a calibrated dial. A simpler version of the principle of expansion employs a metallic coil which produces varying tension on an indicating mechanism in response to temperature changes.

Thermocouples and *thermistors* indicate temperature electrically. Their accuracy and rapidity of response make them suitable for microclimatic observations in air, soil, plant tissues, clothing, and other specialized exposures. The thermocouple consists of a pair of junctions of two unlike metals. When one junction is kept at a constant temperature and the other is exposed to a different temperature, the electromotive force generated in the circuit can be measured by a potentiometer calibrated in degrees. The thermistor, a semiconducting ceramic element, offers less resistance to the flow of current as its temperature increases. Temperature can thus be indicated as a function of current.

For the purpose of climatic records registering thermometers measure the maximum and minimum temperatures. The simplest *maximum thermometer* is a mercury thermometer with a constriction in the bore near the bulb. (See Figure 2.8.) The constriction allows the expanding

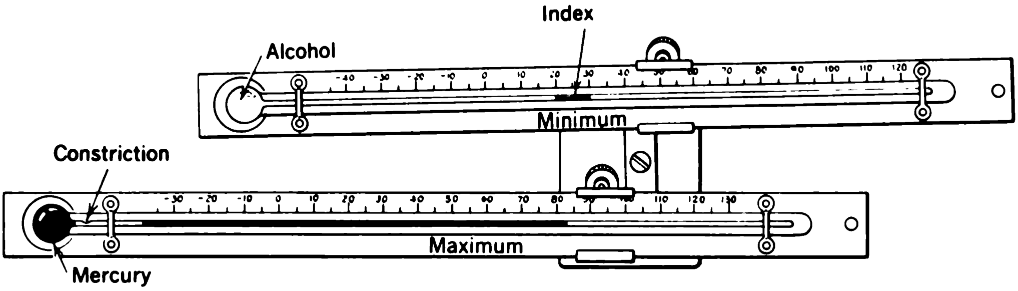


Figure 2.8
The maximum and minimum thermometers.

mercury to pass as the temperature rises, but when cooling occurs the column of mercury breaks at the constriction leaving a part in the bore to register the highest temperature attained. The maximum thermometer is mounted horizontally and is reset by whirling so that centrifugal force pulls the detached thread of mercury down past the constriction. The fever thermometer used by physicians is a specially calibrated maximum thermometer.

The *minimum thermometer* has a larger bore and its fluid is colorless alcohol. A tiny, dark index in the shape of a long dumbbell is placed in the bore below the top of the alcohol column. The minimum thermometer is mounted horizontally and as the alcohol contracts with the decreasing temperature the meniscus (concave surface) of the alcohol pulls the index down. When the meniscus moves up the bore, however, it leaves the index behind to register the lowest temperature. Resetting of the minimum thermometer is accomplished by inverting the stem until the index slides down to the meniscus.

Where a permanent, continuous record of temperature is desired a *thermograph* is used (Figure 2.9). The thermograph consists of an element responsive to temperature changes, a system of levers to translate these changes to a pen arm, and a cylindrical clock drum around which a calibrated chart is mounted. Most thermographs have eight-day clocks and a chart which provides for a continuous ink trace of temperature for seven days. A common temperature element employed is the *Bourdon tube,* a flat, curved tube of phosphor bronze filled with alcohol. The tube changes its curvature in response to temperature changes. Bimetallic elements are also used in thermographs. The readings obtained from a thermograph are not as accurate as those from a mercury thermometer, but frequent checking of the thermograph traces against an accurate thermometer makes a corrected thermograph chart valuable for climatological work.

For most purposes thermometers are read to the nearest one-tenth degree. Temperature differences smaller than one-tenth degree cannot be indicated accurately without special instruments, and the results would be of no added value for most climatic records.

TEMPERATURE RECORDS

For climatological purposes, several kinds of temperature values are desired. Comparatively few stations in the world take hourly temperature readings although an hourly record or a continuous record is highly desirable. From a thermograph trace or from hourly records the daily march of temperature may be determined. Probably the most-used basic temperature value is the *daily mean,* from which monthly and annual values may be computed. The daily mean temperature is found by adding the 24-hour maximum to the 24-hour minimum and dividing by two. (Hourly temperatures during the 24-hour day would provide a better basis for the mean, but at many stations they are not recorded.) The difference between the highest and lowest temperatures of the day is the *diurnal* (daily) *range.* Ordinarily the observations of the maximum and minimum thermometers are the bases for the daily mean tempera-

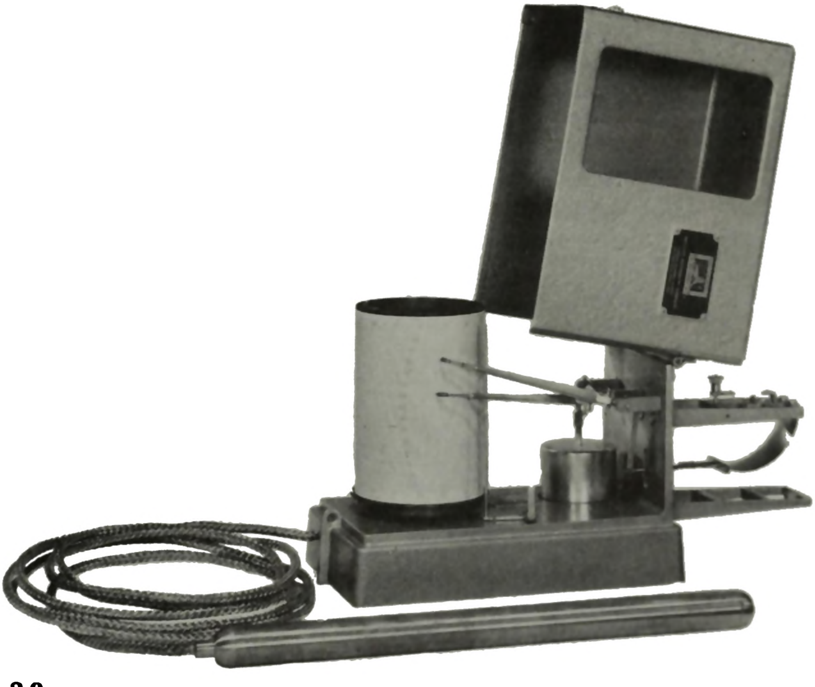


Figure 2.9
Double-recording thermograph. The Bourdon tube on the right activates a pen arm to record air temperature. The element in the foreground translates temperature changes to the second pen arm via a capillary tube from another location in air, water, or soil. (Courtesy Belfort Instrument Company)

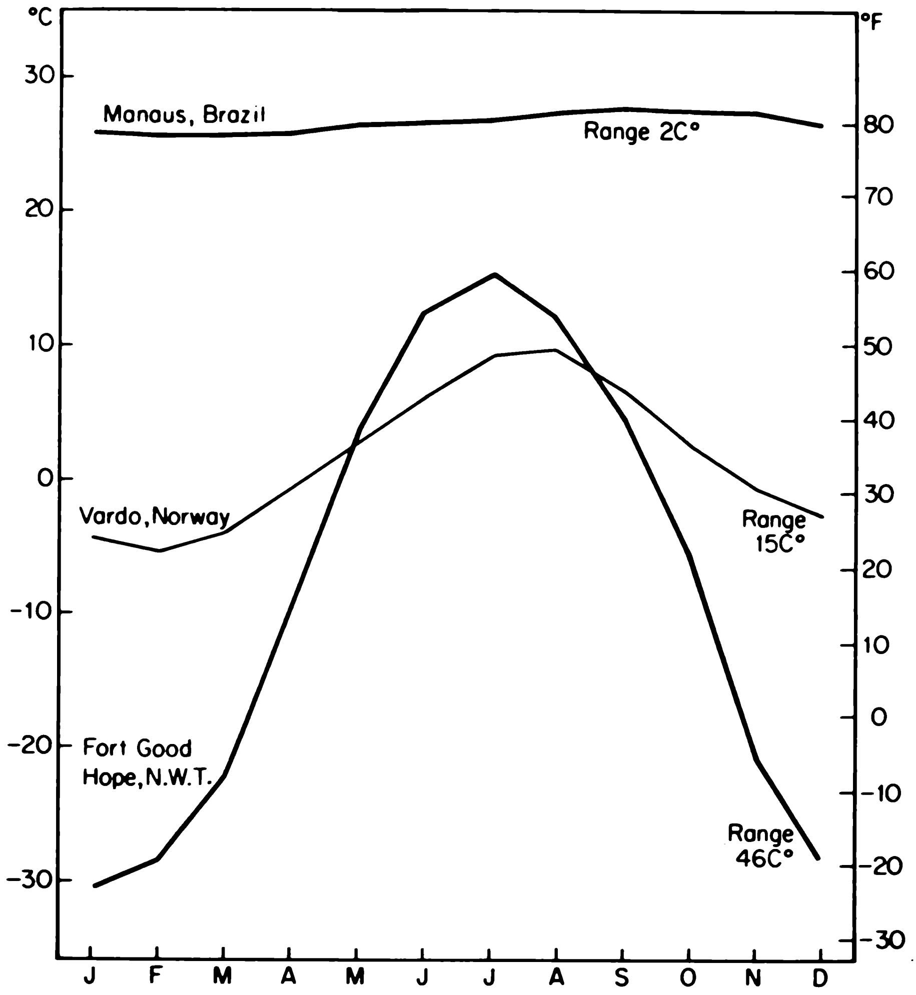


Figure 2.10
Annual variations of mean monthly temperature at selected stations.

ture and the diurnal range. The *mean monthly* temperature is found by adding the daily means and dividing by the number of days in the month. Mean monthly values for the year indicate the *annual march* of temperature through the seasons (see Figure 2.10). The *annual range* is the difference between the mean temperatures of the warmest and coldest months. For most stations in the Northern Hemisphere the warmest month is July and the coldest January. When corresponding temperature values for a number of years are averaged, a generalized value useful in climatic description is obtained. However, such averages tend to obscure the year-to-year variability and climatic changes or cycles.

Especially in the middle and high latitudes the length of the *frost-free season* is commonly a part of the temperature record. Frost-free season is defined as the number of days during which temperatures are continuously above 0°C. The mean frost-free season is the difference in days between the mean date of the last frost in spring and the mean date of the first frost in autumn. It is regarded as an important consideration in agriculture, but there is actually no simple, direct relationship between plant growth and freedom from frost.

HORIZONTAL TEMPERATURE DISTRIBUTION

The differences in temperature from place to place are quite as important as the differences from time to time. An ever-expanding network of temperature-recording stations makes it possible to describe areal differences with increasing accuracy. On maps, the horizontal distribution of temperature is commonly shown by means of *isotherms,* lines connecting points that have equal temperatures. (See Figures 2.11 and 2.12.) On weather maps of small areas the actual observed temperatures are used as a basis for drawing isotherms, but on continental or world maps mean

TABLE 2.3
Records of Extreme Temperatures

Record	*°C*	*°F*	*Location*	*Date*
Highest official air temperature	58	136	Azizia, Libya	Sept. 13, 1922
Highest U.S. temperature	57	134	Greenland Ranch, Death Valley, California	July 10, 1913
Highest mean annual temperature	31	88	Lugh, Somalia	13-year mean
Lowest official temperature in Northern Hemisphere	-68	-90	Verkhoyansk, Siberia	Feb. 5 and 7, 1892
Lowest official temperature in Western Hemisphere	-66	-87	On Greenland ice cap at 2,990 m	Dec. 6, 1949
Lowest temperature on North American continent	-63	-81	Snag, Yukon	Feb. 3, 1947
Lowest U.S. temperature	-62.1	-79.8	Prospect Creek Camp, Alaska	Jan. 23, 1971
Lowest 48-state temperature	-57	-70	Rogers Pass, Mont.	Jan. 20, 1954
Lowest record by U.S. observers	-78.8	-109.8	Amundsen-Scott Station (90° S)	July 14, 1963
Lowest world surface air temperature	-88.3	-126.9	Vostok Soviet Station (78°27′ S 106°52′E at 3,420 m)	Aug. 24, 1960

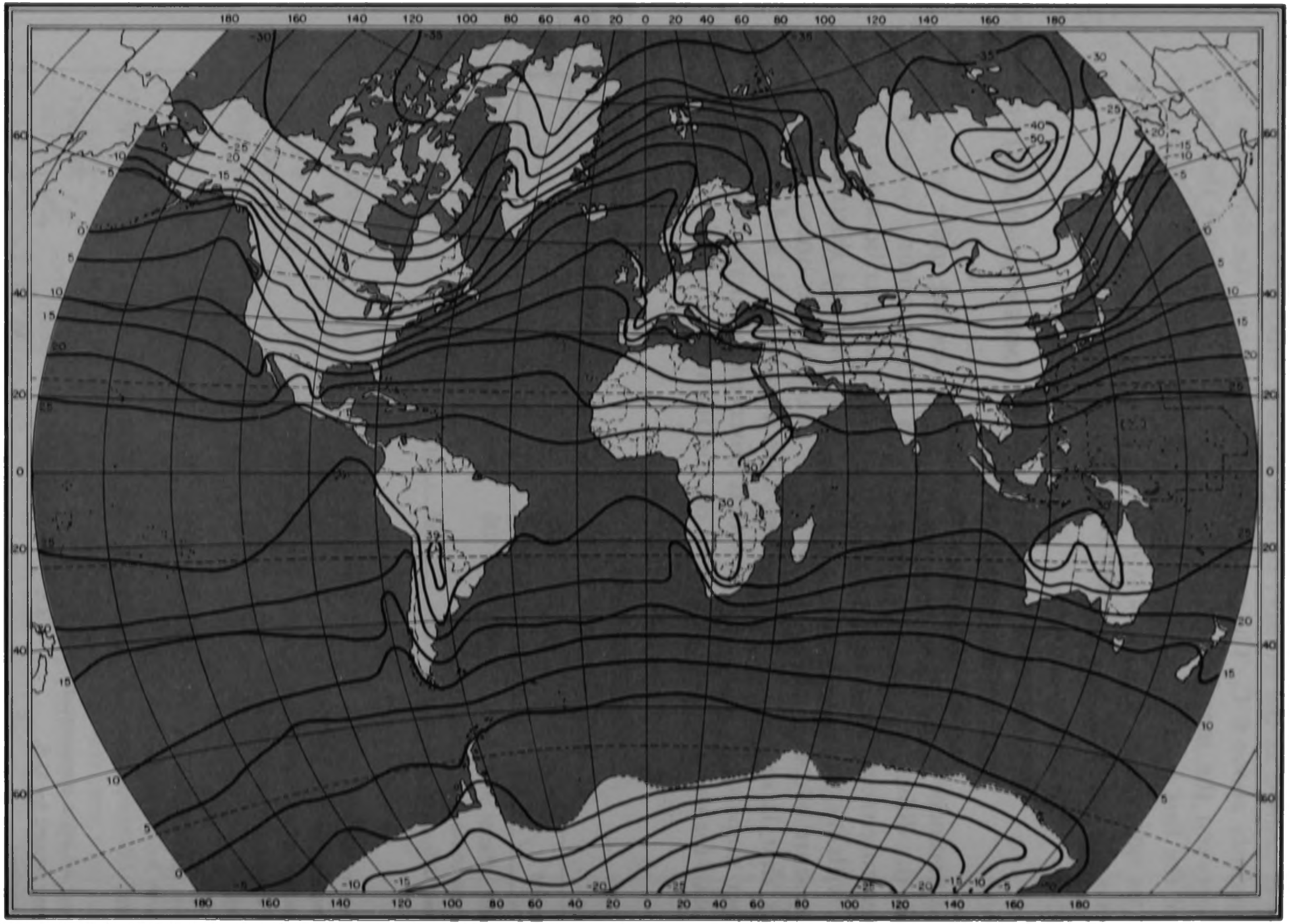


Figure 2.11
World mean sea-level temperatures in January in °C. (modified Van der Grinten projection)

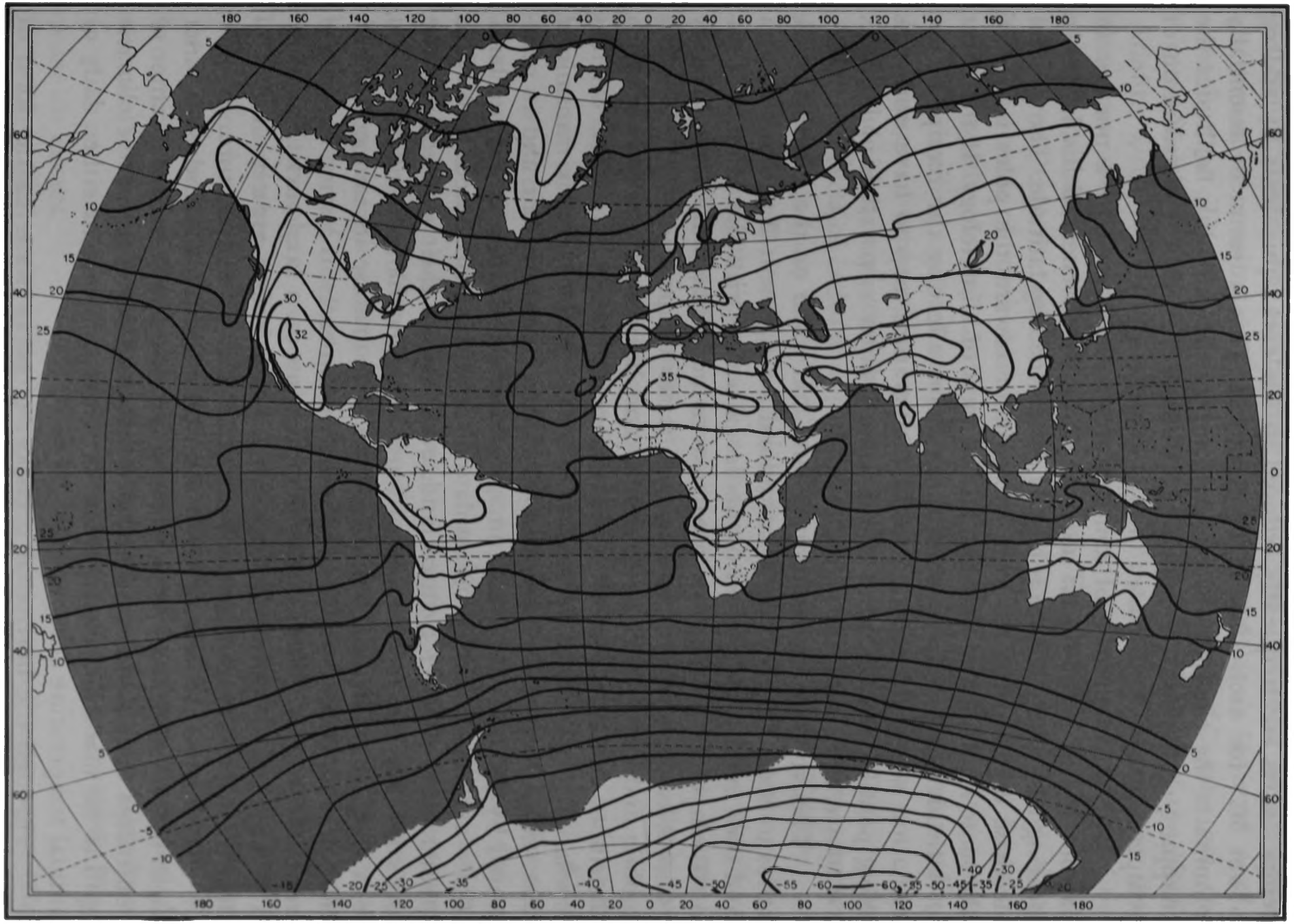


Figure 2.12
World mean sea-level temperatures in July in °C. (modified Van der Grinten projection)

temperatures are frequently reduced to sea-level equivalents by adding about 6C° for each 1000 m of elevation. This adjustment essentially eliminates the effect of altitude on temperature and thus facilitates the mapping of horizontal temperature differences. Vertical distribution of temperature will be discussed in the next section.

The world pattern of temperature is determined by a number of factors. It has already been noted that the effectiveness of insolation in heating the earth's surface is largely determined by latitude. The general decrease in temperatures from the equator toward the poles is one of the most fundamental and best-known facts of climatology. If the effect of latitude were the only controlling factor affecting the receipt of net radiation we would expect a world temperature map to have isotherms lying parallel to each other in the same fashion as parallels of latitude. Of course, such is not the actual case.

The irregular distribution of land and water on the earth's surface tends to break up the orderly latitudinal arrangement. Land areas warm and cool more rapidly than do bodies of water, with the result that annual temperature ranges are greater over land. There are three primary reasons for the contrasts in land and water temperatures. (1) Water is mobile and experiences both vertical and horizontal movements which distribute energy absorbed at the surface throughout its mass, whereas insolation is absorbed by land only at the surface and is transmitted downward slowly by conduction. (2) Water is translucent and is penetrated by radiant energy to a much greater depth than is opaque land. Thus a given quantity of insolation must be distributed through a greater mass of water than of land, even though their surface areas are the same. (3) The *specific heat* of water is higher than that of land. That is, a given mass of water requires more heat to raise its temperature 1° than does an equal mass of dry land. Consequently, the same amount of insolation will produce a higher temperature on a land surface than on a water surface. Conversely, in cooling, water loses a greater amount of heat than does land to produce the same drop in temperature.

The general effect of the contrast in heating of land and water areas is to produce cooler winters and warmer summers in the center of continents than along coasts and over oceans. Coastal or marine climates tend to be moderate, experiencing no great extremes in either daily or annual temperature changes.

Through the horizontal transport of ocean water in the form of currents and drifts heat is carried from one part of the earth to another. Thus, an ocean current travelling poleward warms overlying air, producing air temperatures higher than would normally be expected for the latitude. A current moving toward the equator will produce cooler air temperatures. An outstanding example of the effect of an ocean current upon temperature is that of the North Atlantic Drift off northwest Europe. The January isotherms for −5° and 0°C intercept the United States' east coast in the vicinity of latitude 40° N, but the North Atlantic Drift carries warm water into the northeast Atlantic so that these two isotherms strike the coasts of Great Britain and the Scandinavian Peninsula at a much higher latitude. Examples of the effect of a cold ocean current can be seen in the temperature distribution along the west coasts of South America and Africa. The Peru (South America) and Benguela (Africa) Currents flow northward along the west coasts, where upwelling causes temperatures to be lower than in corresponding latitudes on east sides of the continents.

The horizontal distribution of sea-surface temperatures results from the effects of latitude and the seasons as well as of ocean currents. Extremes range from below 0°C in high latitudes to about 30°C in certain tropical gulfs. Annual variations in temperature of the sea surface are greatest along the east coasts in the middle latitudes of the Northern Hemisphere, the extremes occurring in February and August. There is thus a lag of about two months in heating and cooling of the ocean surfaces behind the periods of high and low sun.

It is obvious that neither oceans nor ocean currents can have their maximum effect upon temperature unless the prevailing winds blow from the water to the land. Prevailing wind direction and the movements of air masses have a direct influence upon the average temperatures of an area. This is well illustrated by the winter flow of air across the United States. The west coast experiences the relatively warm temperatures characteristic of air blown from the Pacific Ocean. The eastern part of the nation receives cooler air from the continent in the general west-to-east movement of storms and air masses and the immediate effect of the Atlantic Ocean upon east coast temperatures is somewhat restricted.

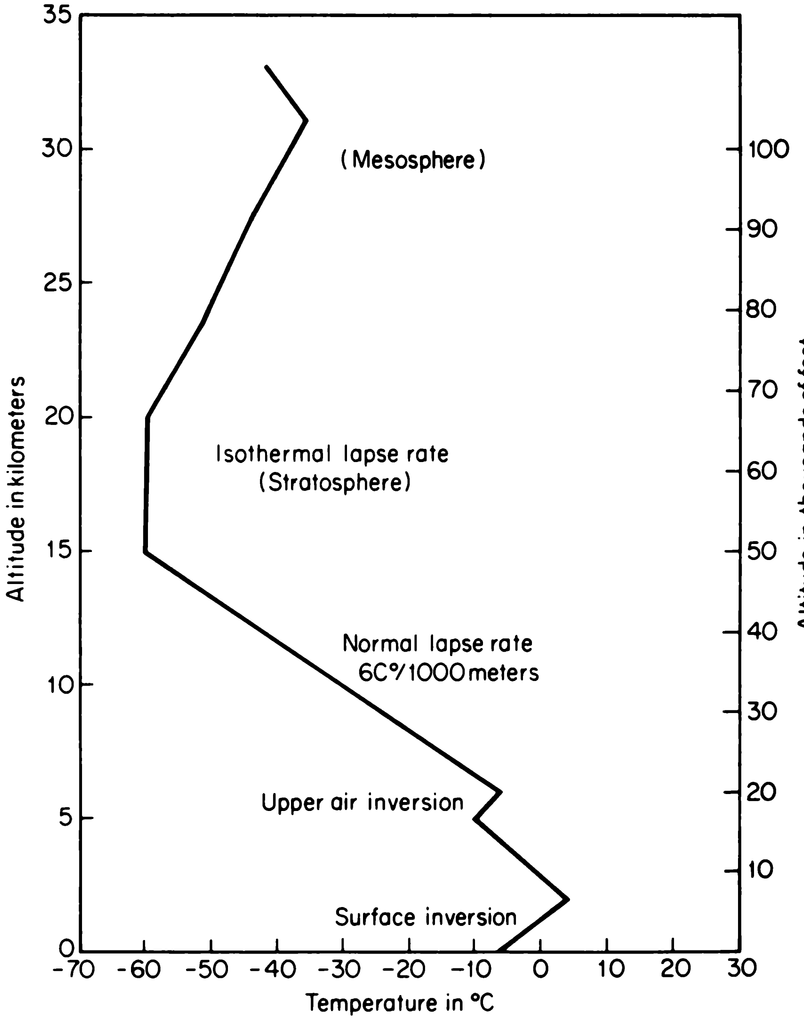
Still another influence on horizontal temperature distribution is mountain barriers. The barrier effect of mountain ranges tends to restrict the movement of cold air masses. In the United States the Rocky Mountain barrier assists the prevailing westerlies in diverting most of the cold outbursts from Canada to the east. Similarly, the Himalayas in Asia and the Alps in Europe protect the regions to the south from polar air masses.

On a local scale, topographic relief exerts an influence upon temperatures. In the Northern Hemisphere north-facing slopes generally receive less insolation than south-facing slopes, and temperatures are normally lower. Drainage of cold air into valleys at night also affects local temperature distribution.

VERTICAL DISTRIBUTION OF TEMPERATURE

The permanent snow caps on high mountains, even in the tropics, indicate the decrease of temperature with altitude. Observations of the temperature in the upper air by means of instruments attached to kites, balloons, airplanes, and rockets have shown that there is a fairly regular decrease in temperature with an increase in altitude. This condition extends upward to the tropopause. The average rate of temperature decrease upward in the troposphere is about 6C° for each 1,000 m of altitude, or about 3.5F° per 1,000 ft. This vertical gradient of temperature is commonly referred to as the *normal lapse rate,* but it varies with height, season, latitude, and other factors. The normal lapse rate represents the average of many observations of vertical temperature distribution and should not be confused with the *actual lapse rate,* which indicates temperature values above a given location at a given time. Indeed,

Figure 2.13
Lapse rate components of a hypothetical vertical temperature sounding.



the actual lapse rate of temperature does not necessarily show a decrease with altitude. Where observations show no change with altitude, the lapse rate is termed *isothermal.* Such a condition never occurs over a very great range of elevation nor for a long period of time. (See Figure 2.13.)

Certain special conditions in the lower troposphere may produce a reversal of the normal rate so that the temperature actually increases with an increase in altitude. This is known as a *temperature inversion,* and the lapse rate is said to be inverted. Temperature inversions near the earth's surface may be produced in five ways. (1) When the earth's surface radiates more heat than it absorbs, as on a clear night or at high latitudes in winter, it cools and consequently lowers the temperature of the adjacent layer of air. Inversions due to radiation often are well developed over snow surfaces. They are not likely to occur over water, which is slow to cool, unless the water is already cooler than the overlying air. (2) Because of its greater density, cold air from hilltops and slopes tends to collect in valley bottoms, creating an inverted lapse rate up the slopes as well as in the free air over the valley floor. Air drainage inversions are frequently associated with spring frosts in middle latitudes, and for this reason fruit growers prefer gentle slopes to valley bottoms for orchard sites. (3) When two air masses with different temperature characteristics come together, the colder air, being more dense, tends to push underneath the warmer air and replace it. The boundary zones along which two air masses meet are called *fronts,* and the inverted lapse rate which results is a *frontal inversion.* Frontal inversions are not confined to the lower layers of the troposphere as are the types mentioned previously; they may form at upper levels wherever cold air underruns warm air or warm air advances above cold air. (4) *Advection* of warm air over a cold surface creates an inversion in the lower layers of the air mass as the warm air is cooled by conduction. This process is common when warm air passes over a cold water surface, cold land, or snowfields. (5) Another type of inversion, the *subsidence inversion,* forms in an air mass when a large body of air subsides and spreads out above a lower layer. In the process the air heats dynamically more in the upper portion than at its base. Inversions of this type may develop at considerable altitudes. They are especially common in the region of the trade winds.

Additional Readings

BUDYKO, M. I., *The Heat Balance of the Earth's Surface,* translated by Nina Stepanova. Washington: U.S. Department of Commerce, Weather Bureau, 1958.

DAVIS, P. A., "An Analysis of the Atmospheric Heat Budget," *Journ. Atmos. Sci.,* 20 (1963), 5–22.

GATES, DAVID M., "Spectral Distribution of Solar Radiation at the Earth's Surface," *Science,* 151, 3710 (1966), 523–29.

HOUGHTON, HENRY G., "On the Annual Heat Balance of the Northern Hemisphere," *Journ. of Met.,* 11 (1954), 1–9.

KONDRATYEV, K. YA., AND G. A. NIKOISKY, "Solar Radiation and Solar Activity," *Quart. Journ. Roy. Met. Soc.,* 96, 409 (1970), 509–22.

LUDLUM, DAVID M., "Extremes of Heat in the United States," *Weatherwise,* 16, 3 (1963), 108–29.

———, "Extremes of Cold in the United States," *Weatherwise,* 16, 6 (1963), 275–91.

MILLER, DAVID H., "The Heat and Water Budget of the Earth's Surface," *Adv. in Geophys.,* 11 (1965), 175–302.

SUTTON, SIR GRAHAM, "Scales of Temperature," *Weather,* 18, 5 (1963), 130–34.

chapter three

Atmospheric Moisture

1793985

Moisture in the atmosphere plays such a significant role in weather and climate that it is commonly treated separately from the other constituents of air. In one or more of its forms atmospheric moisture is a factor in humidity, cloudiness, precipitation, and visibility. Water vapor and clouds affect the transmission of radiation both to and from the earth's surface. Through the process of evaporation water vapor becomes an important medium for conveying latent heat into the air, thus giving it a function in the heat exchange as well as in the moisture exchange between the earth and the atmosphere. Atmospheric water is gained by evaporation but lost by precipitation after complex intervening processes of horizontal and vertical transport and changes in physical state. Only a minute fraction of the earth's water is stored as clouds and vapor in the atmosphere at any one time. The net amount at the end of any given period for a particular column of air is an algebraic summation of: the amount stored from a previous period, the gain by evaporation, the loss by precipitation, and the gain or loss by horizontal transport. This relationship expresses the *water budget of the atmosphere.*

WATER VAPOR

Humidity is a general term connoting the amount of water vapor in the air. *Specific humidity* is the ratio of the mass of water vapor actually in the air to a unit mass of air. Thus a kilogram (1,000 g) of air of which 12 g are water vapor has a specific humidity of 12 g per kg. A closely related term is *mixing ratio,* which is the mass of water vapor per unit mass of dry air. The mixing ratio is, in a sense, a recipe for an admixture of water vapor and dry air. A value of 12 g per kg would entail a total of 1,012 g for the mixture. For most conditions, the specific humidity and the mixing ratio differ insignificantly. Another humidity expression based upon separate consideration of air and water vapor is *vapor pressure.* This is the partial pressure exerted by water vapor in the air, and it is independent of the other gases. It is expressed in the same units used for barometric pressure, that is, millibars or inches of mercury.

When air contains all the water vapor it can hold at a given temperature it is *saturated,* and its actual vapor pressure equals its *saturation vapor pressure.* The air is then at its *dew point* temperature.

A term representing the density of water vapor in the air is *absolute humidity,* which is the mass of water vapor contained in a unit volume of air and is expressed as grams per cubic meter (or in grains per cubic foot). It has wide application in engineering and air conditioning, but it is not often used in meteorology.

Probably the best-known and most-used reference to water vapor is

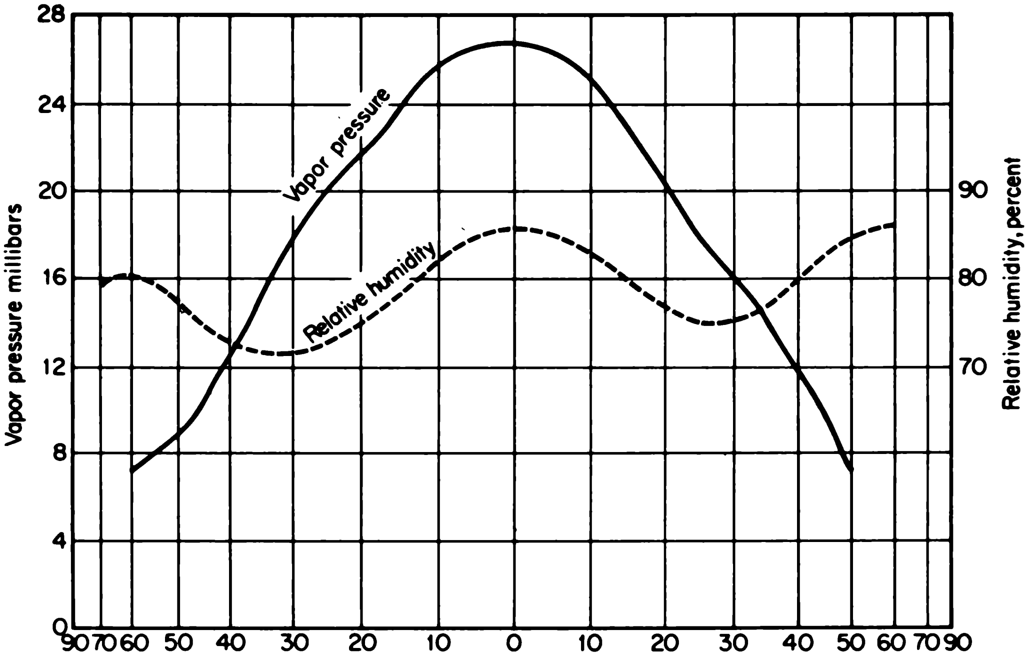


Figure 3.1

Approximate mean latitudinal distribution of relative humidity and vapor pressure.

relative humidity, which is the ratio of the amount of water vapor actually in the air to the amount the air could hold at a given temperature and pressure, that is, the ratio of the actual to the saturated vapor pressure. Thus, if a kilogram of air at constant pressure could hold 12 g of water vapor at a certain temperature but contains only 9 g at that temperature, it has a relative humidity of 75 percent. When the temperature of air is increased, the capacity to hold moisture increases. If no further moisture is added, the result will be a decrease in relative humidity. Conversely, when air temperature is decreased, its capacity for moisture decreases and its relative humidity increases. The average diurnal maximum relative humidity occurs in the early morning hours and the minimum in the early afternoon. Relative humidities tend to be greater over land in winter, except where there are strong summer incursions of moist air, for example, in regions affected by the monsoon. Over the oceans, relative humidity reaches a slight maximum in summer. On mid-latitude mountains there is also a tendency to summer maximums because of strong convective flow of moist air. Figure 3.1 shows the approximate mean latitudinal distribution of relative humidity. Although the actual pattern is modified by other climatic factors, maximums generally prevail near the equator and minimums in the belts of high pressure at subtropical latitudes. Poleward, toward the westerly wind belts, there is again an increase owing to decreasing temperatures. This zonal distribution of relative humidity contrasts somewhat with the average pattern of vapor pressure or specific humidity. Even the desert air of the tropics holds considerable water vapor, although its *relative* humidity is low. In the polar regions low mean temperatures produce low specific humidities.

HUMIDITY MEASUREMENTS

Direct measurement of the actual amount of water vapor in the air is not feasible for ordinary observations. Instead the various humidity values are determined indirectly from the *psychrometer* (Figure 3.2). This instrument is simply two thermometers mounted on the same backing. One is mounted a little lower than the other and has its bulb covered with a piece of muslin or wicking which can be wetted for the observation. It is known as the *wet-bulb thermometer* and the other is the *dry-bulb*. When the psychrometer is swung freely in the air or is aerated by a fan, the loss of heat required to evaporate water from the wet-bulb will cause it to show a lower temperature reading. The difference between the readings is called the *wet-bulb depression*. When the dry-bulb temperature, the wet-bulb depression, and the atmospheric pressure are known any of the standard expressions of humidity can be determined

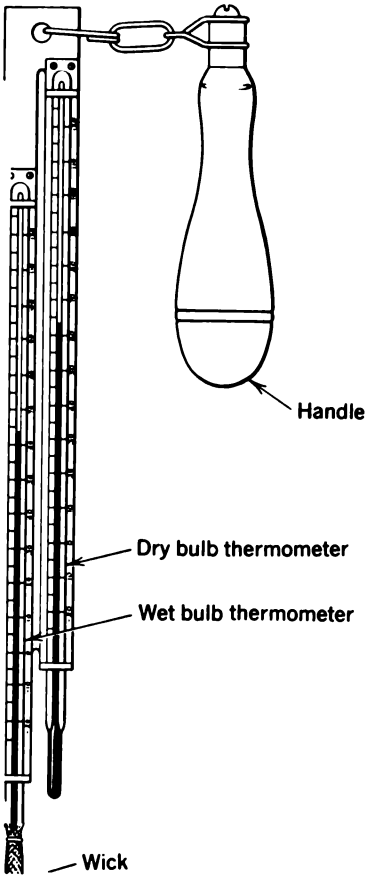


Figure 3.2
Sling psychrometer.

from a series of psychrometric tables, which have been computed by the use of formulas based on mathematical relationships. If there is no depression of the wet-bulb, the air is saturated and the relative humidity is 100 percent.

Relative humidity is one of the humidity values for which there is an instrument that makes a direct measurement. The *hair hygrometer* operates on the principle that human hair lengthens as relative humidity increases and contracts with decreasing relative humidity. The tension of several strands of hair is linked to an indicator. Unfortunately there is such considerable lag in the response, especially at low temperatures, that the hair hygrometer is a much less accurate instrument than the psychrometer. For meteorological purposes, its principle is employed primarily in the *hygrograph,* a recording hygrometer which has a clock-drum and pen arrangement just as in the thermograph.

A device known as the *infrared hygrometer* employs a beam of light projected through the air to a photoelectric detector. Two separate wave lengths of light are used; one is absorbed by water vapor and the other passes through undiminished. The ratio of the infrared light transmitted by the different wave lengths is a direct indication of the amount of water vapor in the light path.

Hygrometers in the home are ordinarily of the hair type. Some "hygrometers" are actually stationary psychrometers, with the wet-bulb covered by a wick which carries a constant flow of water from a small dish or tube. Such instruments may give fairly accurate humidity measurements indoors if the air is calm and proper humidity tables are used, but outdoors the variation of air movement about the wet-bulb produces corresponding variations in the reliability of derived humidity values. A novel humidity "instrument" is the weather house. When the old woman comes out it is supposed to rain; when the young girl appears fair weather is due. One or two strands of catgut, their tendency to twist varying with relative humidity, activate this device.

A type of humidity gauge useful where electrical transmission of the variation is desirable depends upon the fact that the passage of electrical current across a chemically coated strip of plastic is proportional to the amount of moisture adsorbed at its surface. This type of humidity element is in common use in remote-indicating instruments such as the radiosondes used for upper air observations. Still other devices known as *dew point hygrometers* employ artificially cooled surfaces to determine the dew point temperature directly.

PHYSICAL CHANGES OF STATE OF WATER

The chemical compound H_2O occurs naturally in the atmosphere in all three physical states—gas, liquid, and solid—and it may change from one state to any other, always with heat involved (Figure 3.3).

Evaporation, the change of state from liquid to vapor, results when molecules escape from any water surface, whether it be the surface of water bodies, droplets in clouds or fog, or thin films on solids such as soil

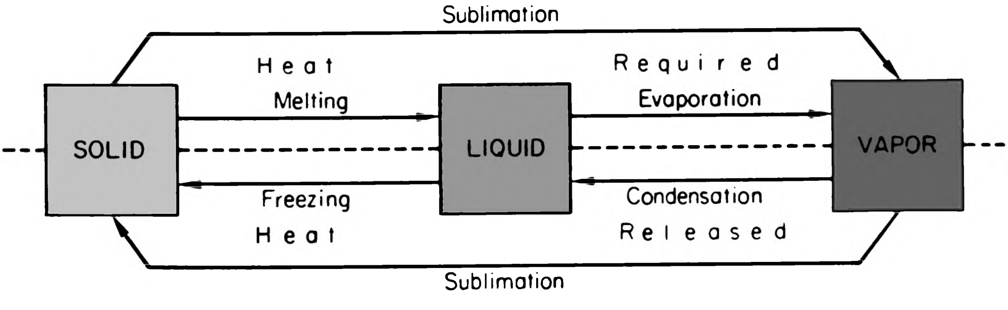


Figure 3.3
Physical changes of state of water.

particles. The process requires energy and is the means by which vast amounts of latent heat are transferred from the earth's surface to the atmosphere. Its rate is dependent upon three major factors: vapor pressure, temperature, and air movement. Evaporation increases as the saturation vapor pressure at the water surface becomes greater than the actual vapor pressure of the adjacent air. It therefore takes place more rapidly into dry air than into air with a high relative humidity. The rate of evaporation also increases with rising water temperature, other factors being equal. When the temperature of the water is higher than that of the air, evaporation always takes place. Finally, wind movement and turbulence replace air near the water surface with less moist air and increase evaporation. The heat used in carrying out evaporation is retained by the water vapor as the *latent heat of vaporization* and does not raise the temperature of the vapor. The latent heat of vaporization ranges in value from nearly 600 calories per gram (cal per g) of water at O°C to about 540 cal at 100°C because as temperature increases the difference between the kinetic energy of the escaping molecules and the average kinetic energy of all the water molecules decreases. The process of evaporation therefore becomes less selective and requires less heat as a growing proportion of the liquid molecules gain the necessary energy to escape from the water surface.

A special case of evaporation is *transpiration,* which entails a loss of water from leaf and stem tissues of growing vegetation. The combined losses of moisture by evaporation and transpiration from a given area are termed *evapotranspiration.*

Measurement of the amount of moisture which air will take up by evaporation is commonly accomplished by recording the water lost from an exposed pan or tank, making due allowance for water added by precipitation. Other types of *evaporimeters* known as *atmometers* employ constantly wetted porous substances that can be exposed to the free air.

More complex measurements to determine the actual amount of water evaporated from soil surfaces and transpired from vegetation prove difficult in practice. Indirect approaches are based on the water balance equation:

$$ET = P - (R + dS)$$

where ET is the loss by evapotranspiration, P the gain by precipitation, R the loss to runoff, and dS is the gain or loss of water stored in the soil. Thus, if precipitation, runoff, and soil storage can be measured or estimated adequately, we can calculate evapotranspiration as a difference.

Since evapotranspiration entails an exchange of energy as well as

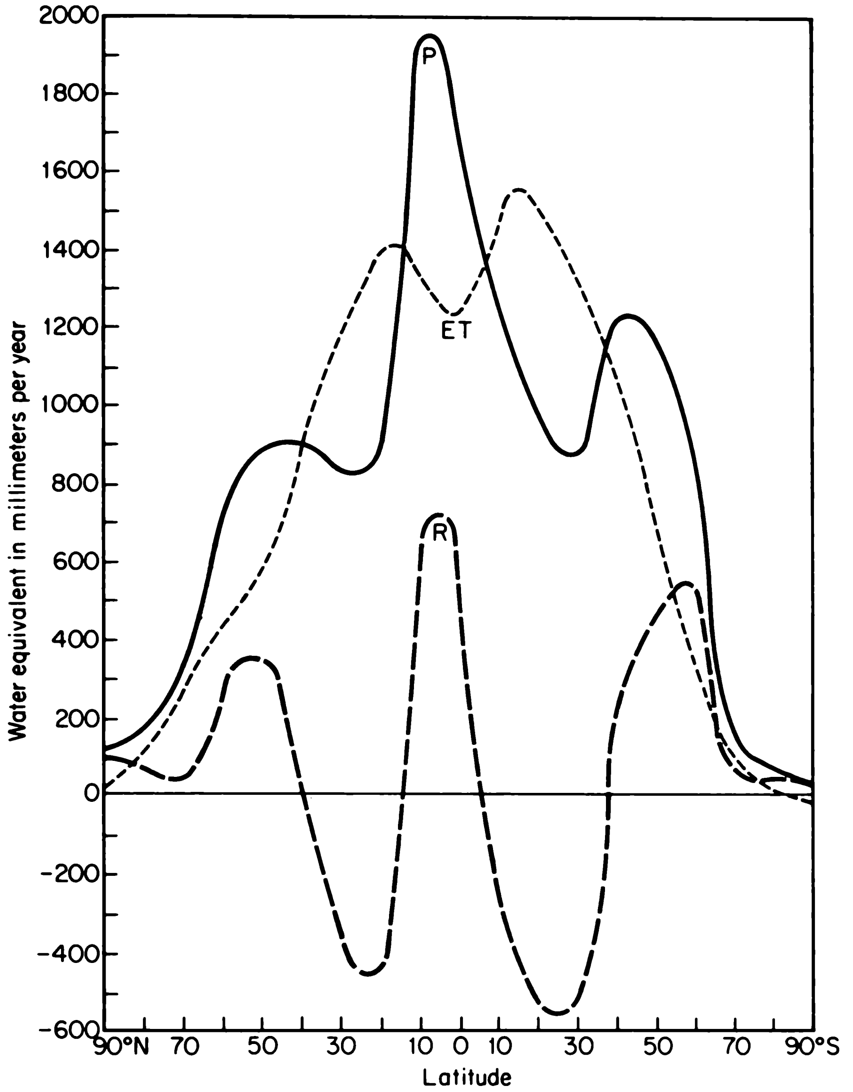


Figure 3.4
Mean annual latitudinal distribution of water budget elements. ET, *evapotranspiration;* P, *precipitation;* R, *runoff.* (After Sellers)

of moisture between the surface and the overlying air, attempts have been made to calculate its value in terms of heat, radiation, temperature, and air movement.

The importance of solar energy in the evaporation process is evident in the latitudinal distribution of evaporation (Figure 3.4). Mean annual evaporation rates are greatest in the vicinity of 20° north and south, where net radiation is at a maximum under the generally clear skies and subsiding air masses of the tropical deserts. The greater evaporation in the Southern Hemisphere results from the larger proportion of ocean surface south of the equator.

Condensation is the change of state from water vapor to liquid water. When moist air comes in contact with cool surfaces, it may be cooled to the point where its capacity to hold water vapor is exceeded by the actual amount in the air. A part of the vapor then condenses into a liquid form on the cool surface, producing *dew.* When this happens the latent heat of vaporization, in this process called the *latent heat of condensation,* is released. Liberation of the latent heat of condensation always tends to slow down the cooling process. It provides one of the main sources of heat energy for atmospheric processes. The temperature at which air becomes saturated upon cooling and at which condensation normally begins is the *condensation temperature* or *dew point.* However, the dew point temperature is not a definite boundary at which all condensation takes place. Condensation also results from cooling in the free air, but it requires the presence of very small *condensation nuclei.* The most active nuclei are hygroscopic particles, which have an affinity for water, and upon which condensation may take place even at temperatures above the dew point, that is, when the relative humidity is less than 100 percent. If sufficient water droplets are formed in this way to become visible, *haze* will result, becoming fog or possibly clouds as the number and density of droplets increase. Among the hygroscopic materials in the atmosphere are salts liberated from bursting bubbles in the foam of ocean spray and numerous chemical compounds in industrial smokes. Insoluble dust particles are less effective nuclei, although they may have greater significance in the formation of ice crystals at high levels.

As cooling and condensation proceed various degrees of *supersaturation* are required to maintain the growth of droplets. The saturation vapor pressure must be considerably exceeded in the air surrounding the smaller droplets and those with nonhygroscopic nuclei. Larger droplets and those having soluble nuclei gain increments of condensation more readily. Laboratory experiments have shown that in carefully washed and filtered air supersaturation as great as 700 percent is required before the beginning of condensation. Since there is normally an abundance of condensation nuclei in the lower atmosphere, visible condensation products can be expected to form whenever the relative humidity reaches 100 percent.

Superficially, the processes of *melting* and *freezing* appear to be remarkably uncomplicated. Ice melts at a temperature of 0°C (32°F) when warmed, but liquid water does not always freeze at that temperature, contrary to popular belief. Foreign material dissolved in the water, suspended in it, or covering it, the amount of water involved in the process, and electrical charges all may lower the freezing point. In free air, small water droplets are known to exist at temperatures far below the "normal" freezing point. This is especially true at upper levels. The supercooled condition can apparently be discontinued by lowering the temperature below a critical point of about −40°C (also −40°F) or by introducing artificial ice nuclei. In the rainmaking experiments of recent years dry ice has been used to produce low temperatures. Silver iodide crystals introduced into a supercooled cloud result in formation of ice particles when temperatures are below −4°C, being most effective at about −15°C. The smaller the droplets of water the lower the temperature at which they will freeze spontaneously. The processes both of freezing and melting involve the *latent heat of fusion* (80 cal per g at 0°C), which is absorbed when ice melts and released when water freezes.

At temperatures below freezing, water may bypass the liquid form in its change of state. When dry air with a temperature well below freezing comes in contact with ice, molecules of the ice (H_2O) pass directly into the vapor state by means of the process of *sublimation*. Sublimation frequently removes snow from the ground in dry winter weather. The reverse process is also known as sublimation, in which both the heat of vaporization and the heat of fusion are released. Sublimation nuclei must be present in free air to produce ice fog or ice clouds. If sublimation occurs on a cold surface the visible product is *frost*, also known as *hoar frost*. Sublimation nuclei probably play an important part in dispersing clouds of supercooled water droplets. A leading theory is that they initiate the formation of small ice crystals from the vapor in the air adjacent to the supercooled droplets. A part of the supercooled water then evaporates into the drier air and the resulting vapor in turn sublimates on the ice crystals. Thus the supercooled droplets do not freeze directly on the crystals, as might be supposed.

PROCESSES OF COOLING TO PRODUCE CONDENSATION AND SUBLIMATION

Some type of cooling process normally is required to sustain condensation or sublimation. The cooling processes are as follows.

1. Adiabatic processes
   - A. A decrease in barometric pressure at the surface, in which case fog might form
   - B. Rising air, which in turn results from
     - (a) convection
     - (b) convergence of wind currents or air masses (as along fronts)
     - (c) orographic lifting
2. Diabatic process
   - A. Loss of heat by radiation. Direct radiation from moist air may produce fog or clouds.

B. Contact with a cold surface (conduction). Dew, frost, or fog are the normal products. The process may be associated with the movement of air across a cold surface (advection).
C. Mixing with colder air. If the mixture has a temperature below its dew point clouds or fog will form, assuming sufficient effective nuclei are present.

Of all the types of cooling which air may experience, those due to lifting are by far the most important in condensation in free air. When air rises for any reason there is less pressure upon it at the new level and it expands and cools. Upon subsiding it undergoes an increase in pressure and is warmed. These temperature changes, occurring without any heat being added to or subtracted from the air, are termed *adiabatic.* They result from changes in pressure which in turn alter the kinetic energy of the air. *Diabatic,* or nonadiabatic, processes involve the gain or loss of heat by the air from or to an outside source. The rate at which unsaturated air cools as it rises is 10C° per 1,000 m or 5.5F° per 1,000 ft. If rising and cooling continue until condensation begins, however, the latent heat of condensation released into the air reduces the rate of cooling. The new rate of cooling is called the *wet,* or *pseudoadiabatic, rate;* it is the *true,* or *dry adiabatic, rate* modified by the latent heat of condensation. Its value varies with temperature and pressure (and therefore with heights reached by ascending air), averaging about 6C° per 1,000 m or 3.2F° per 1,000 ft.

Adiabatic cooling due to lifting is accomplished by any of the three ways outlined above or by any combination of them. Convection has been discussed previously as a method of heat transfer. As a process leading to adiabatic cooling it is common whenever the earth's surface is warmer than the air above. Convergence occurs when winds of different directions or speeds meet one another. In tropical areas convergence frequently involves air currents with similar temperature characteristics. In middle latitudes there is likely to be a considerable difference between the temperatures of converging winds, and the warmer air will ride over the cooler. This special case of convergence is known as *frontal lifting. Orographic lifting* is ascension of air caused or intensified by any topographic obstruction on the earth's surface. Ranges of mountains and hills are most effective in orographic lifting, forcing winds which blow against them to adopt vertical components. Even a slight rise in land elevation, as on a coastal plain, can induce some lifting. Assume that the wind blows in from the sea across a flat coastal plain. The frictional drag of the land is greater than that of the sea, and the lower layers of air will consequently be slowed. The air coming from seaward will be forced to rise over this barrier of congested air in order to proceed inland. Because of the tendency of air to pile up against the barrier, the effect of any surface which induces orographic lifting extends for some distance to windward.

CLOUDS: THEIR FORMATION AND CLASSIFICATION

Clouds are visible aggregates of water droplets, ice particles, or a mixture of both. Most clouds result from cooling due to lifting of air. Those associated with strong rising air currents have vertical development and a puffy appearance, whereas those resulting from gentler lifting or other methods of cooling tend to spread out into layers. Although their method of formation is largely responsible for their appearance, clouds are classified primarily on the basis of their height, shape, color, and transmission or reflection of light. The World Meteorological Organization has established a detailed cloud classification with categories for genera, species, and varieties and with provisions for naming clouds that have undergone changes in form. Table 3.1 indicates the use of Latin names and the major elements of the classification.

It will be sufficient here to divide all clouds of the troposphere into four families: high, middle, low, and clouds with vertical development. Average heights of cloud bases vary with latitude, being lower in polar regions and in winter. High clouds are the *cirrus* types, always com-

Figure 3.5
Cirrus (Cirrus uncinus). (NOAA)

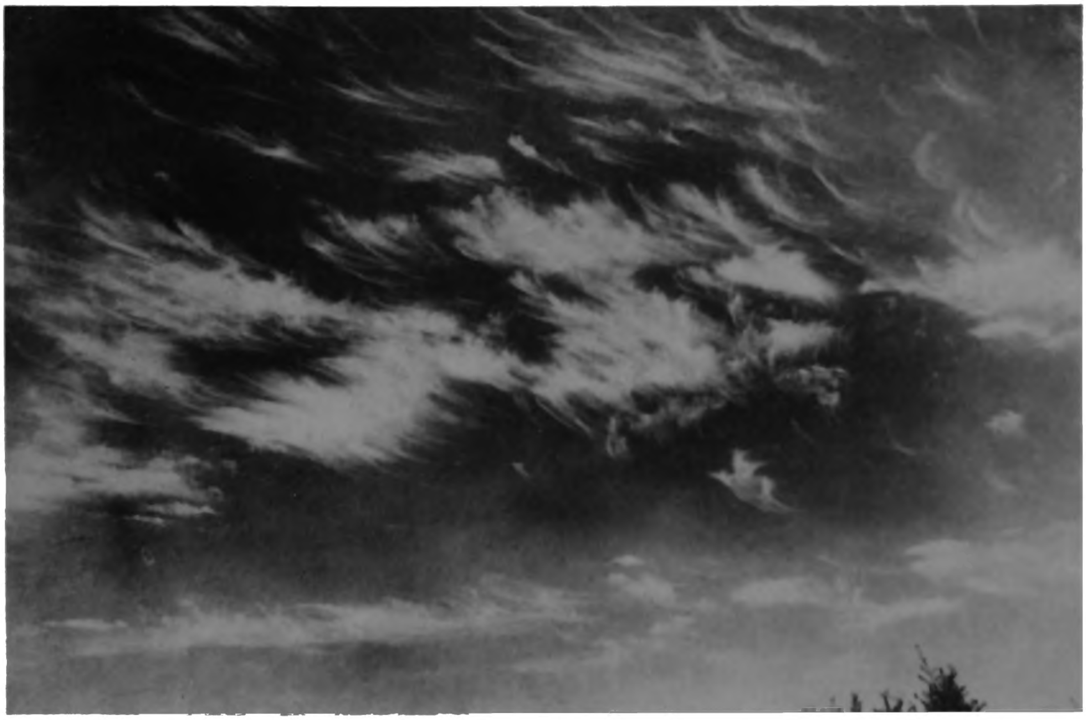


TABLE 3.1
Classification of Clouds*

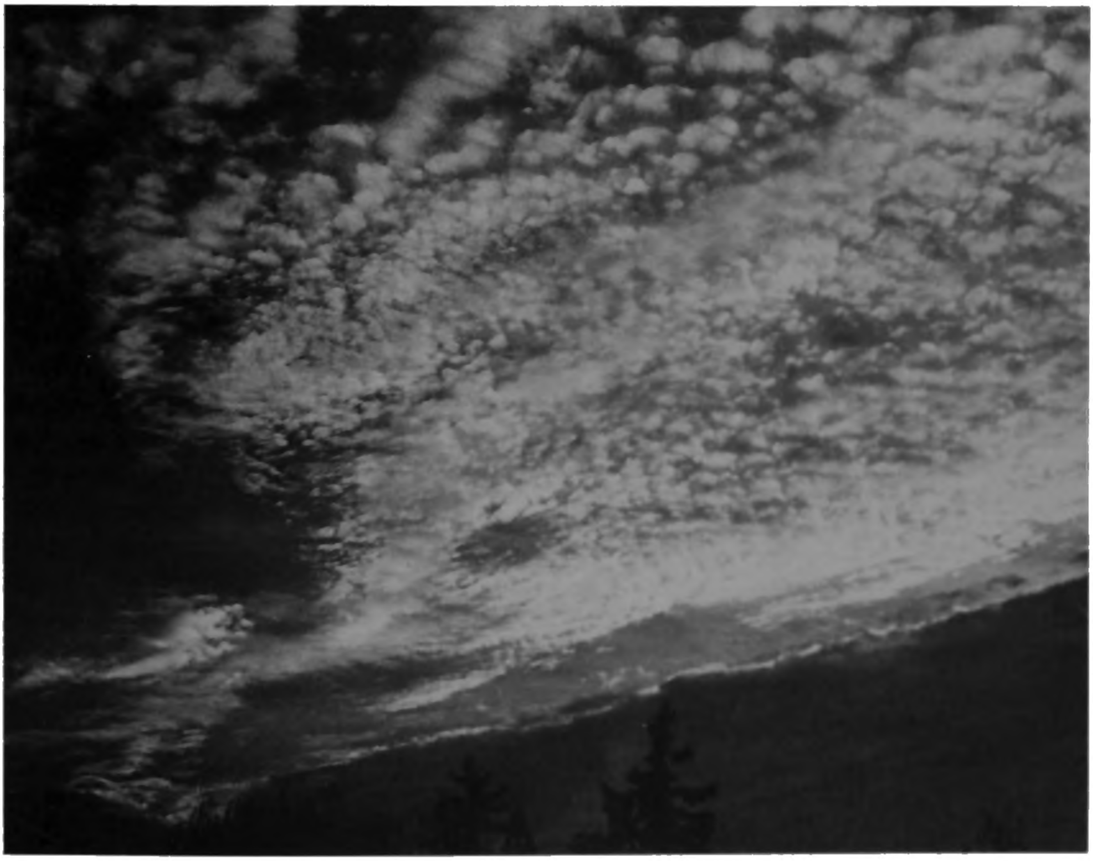
Genera	*Species*	*Varieties*
CIRRUS	fibratus	intortus
	uncinus	radiatus
	spissatus	vertebratus
	castellanus	duplicatus
	floccus	
CIRROCUMULUS	stratiformis	undulatus
	lenticularis	lacunosus
	castellanus	
	floccus	
CIRROSTRATUS	fibratus	duplicatus
	nebulosus	undulatus
ALTOCUMULUS	stratiformis	translucidus
	lenticularis	perlucidus
	castellanus	opacus
	floccus	duplicatus
		undulatus
		radiatus
		lacunosus
ALTOSTRATUS	–	translucidus
		opacus
		duplicatus
		undulatus
		radiatus
NIMBOSTRATUS	–	–
STRATOCUMULUS	stratiformis	translucidus
	lenticularis	perlucidus
	castellanus	opacus
		duplicatus
		undulatus
		radiatus
		lacunosus
STRATUS	nebulosus	opacus
	fractus	translucidus
		undulatus
CUMULUS	humilis	radiatus
	mediocris	
	congestus	
	fractus	
CUMULONIMBUS	calvus	–
	capillatus	

*From *International Cloud Atlas.* Geneva: World Meteorological Organization (1956), p. 7.

posed of ice crystals, whose average heights range from 5 km above the earth's surface to the tropopause. Cirrus is nearly transparent, white, and fibrous or silky (Figure 3.5). *Cirrocumulus* clouds are a cirriform layer or patch of small white flakes or tiny globules arranged in distinct groups or lines. They may have the appearance of ripples similar to sand on a beach. Cirrocumulus is not a common cloud. *Cirrostratus* is a thin white veil of cirrus. It is nearly transparent so that the sun, moon, and brighter stars show through distinctly. Cirrus clouds frequently can be detected by the *halos* which they create around the sun or moon and can thus be distinguished from haze or light fog. Halos result from the refraction of light by ice crystals suspended in the air.

The middle clouds predominate at heights ranging from 2 to 8 km above ground level. They are composed of water droplets, ice crystals, or mixtures of both. The principal types are *altocumulus* and *altostratus*. Altocumulus contains layers or patches of globular clouds, usually arranged in fairly regular patterns of lines, groups, or waves (Figure 3.6). Vertical air currents in the layers occupied by altocumulus may cause the clouds to build upward. Isolated altocumulus may form in the wave of an air stream as it rises over a mountain or a convective

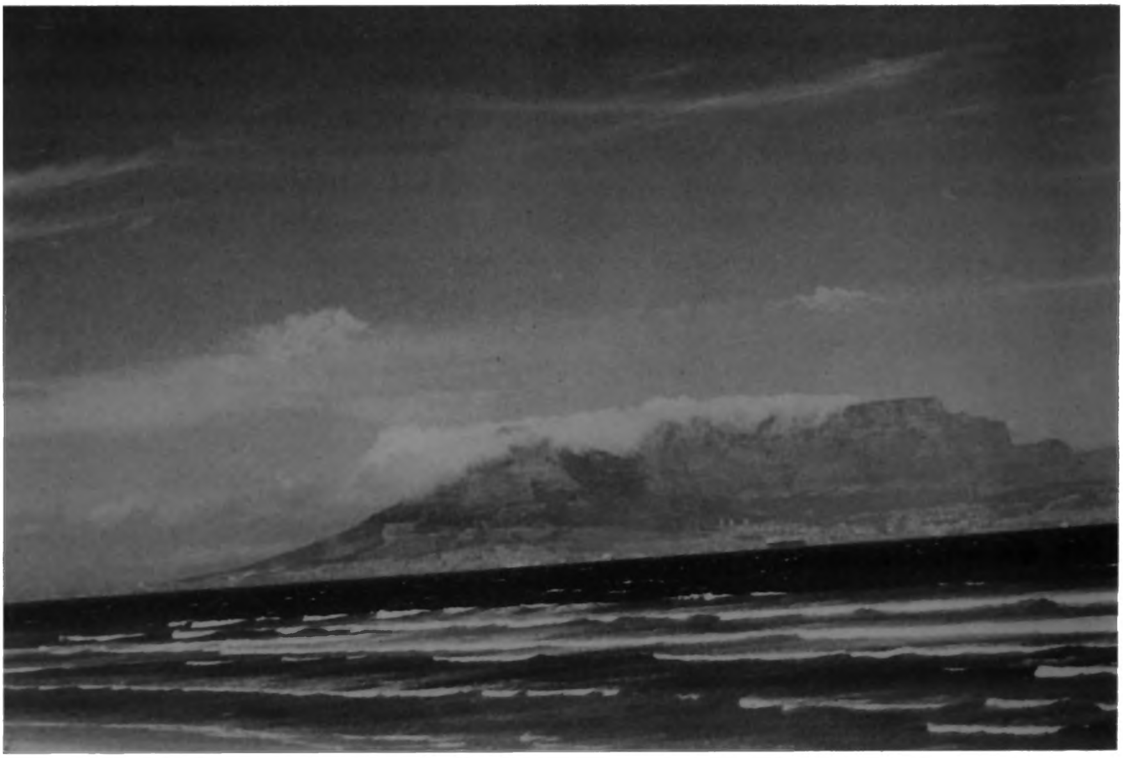
Figure 3.6
Altocumulus (Altocumulus stratiformis). (Photo by author)



column, provided the rise is sufficient to cool the air below its dew point. The banner clouds above peaks (or above local "chimneys" of convection) are of this type. Because of their lens shape they are named *altocumulus lenticularis.* Altostratus is a fibrous veil, gray or blue-gray, and thicker than the higher cirrostratus, although it may merge gradually into the latter. It ordinarily does not exhibit halo phenomena, but another optical effect, the *corona,* sometimes appears as a smaller circle of light around the sun or moon in both altostratus and altocumulus. A halo is produced by refraction in ice crystals, but the corona results from the diffraction of light by water droplets. The corona is smaller than the halo and its diameter is inversely proportional to the size of the cloud droplets. The red of the faint color band is on the outside, that is, away from the sun or moon; in the halo the red is on the inside. It is common for altostratus to change into altocumulus and vice versa. Precipitation may fall from altostratus, or altocumulus, but it does not necessarily reach the ground. Sometimes this *virga* is visible hanging from the bottom of the cloud in dark, sinuous streaks. When the cloud layer lowers to become somewhat thicker and darker and falling rain or snow obscures its base it is called *nimbostratus.*

The base level of low clouds varies from very near the ground to

Figure 3.7
"Table Cloth" on Table Mountain, Cape Town, South Africa. The stratus layer forms as southeasterly winds lift moist air over the mountain. The cloud evaporates in the descending flow on the lee (near) side. (Photo by author)

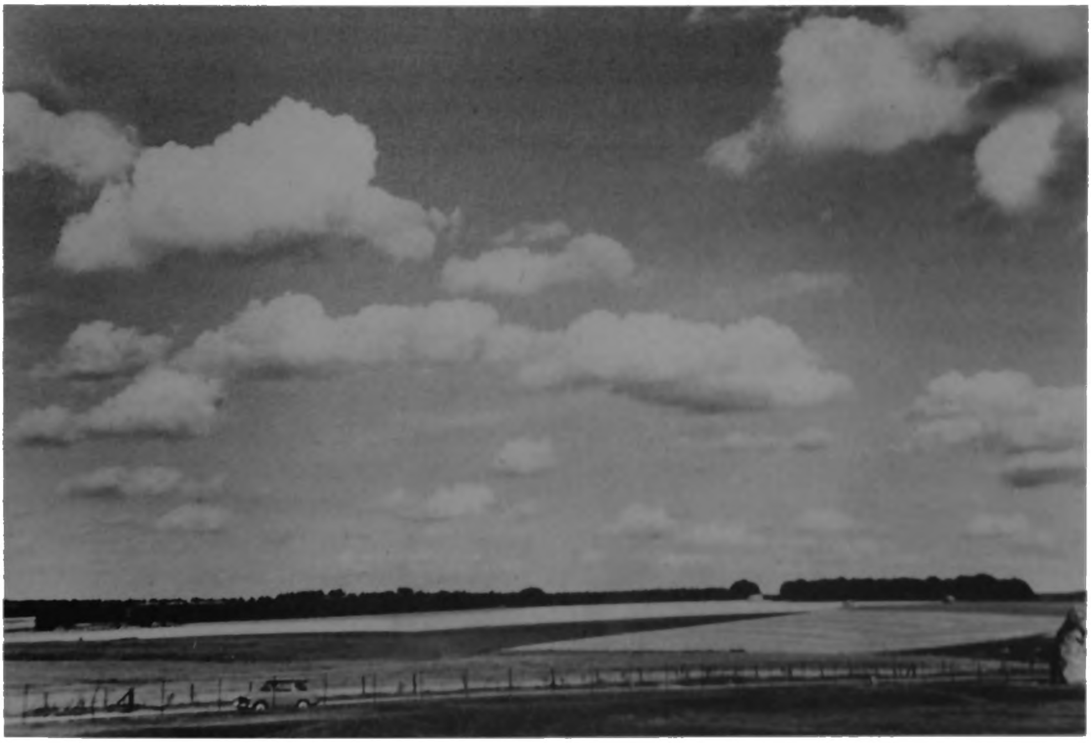


about 2,000 m. The basic type of this family is the *stratus,* a low, uniform layer resembling fog but not resting on the ground. Stratus is frequently formed by the lifting of a fog bank or by the dissipation of its lower layer. If broken up into fragments by the wind it is called *fracto-stratus.* Heat loss by radiation from the top of a stratus layer can cause condensation in the cooled layer above the cloud so that it builds upward. This should not be confused with vertical development due to rising air. When associated with the nimbostratus, stratus sometimes builds downward as well. Rain falling from the upper cloud layer may evaporate on falling through dry air and then recondense in lower saturated layers to form stratus. Precipitation from stratus, if it occurs, is usually light.

Stratocumulus clouds form a low, gray layer composed of globular masses or rolls which are usually arranged in groups, lines, or waves. If the aggregates of thick stratocumulus fuse together completely so that the structure is no longer evident, the clouds are classified as stratus. Note that altocumulus clouds have the same general characteristics. Altocumulus viewed from a mountain or an airplane may be difficult to distinguish from stratocumulus for this reason.

Clouds with vertical development fall into two principal categories:

Figure 3.8
Fair weather cumulus (Cumulus humilis) *on the Salisbury Plain, England. See Figure 3.9 for a satellite view of fair weather cumulus over the same area.* (Photo by author)



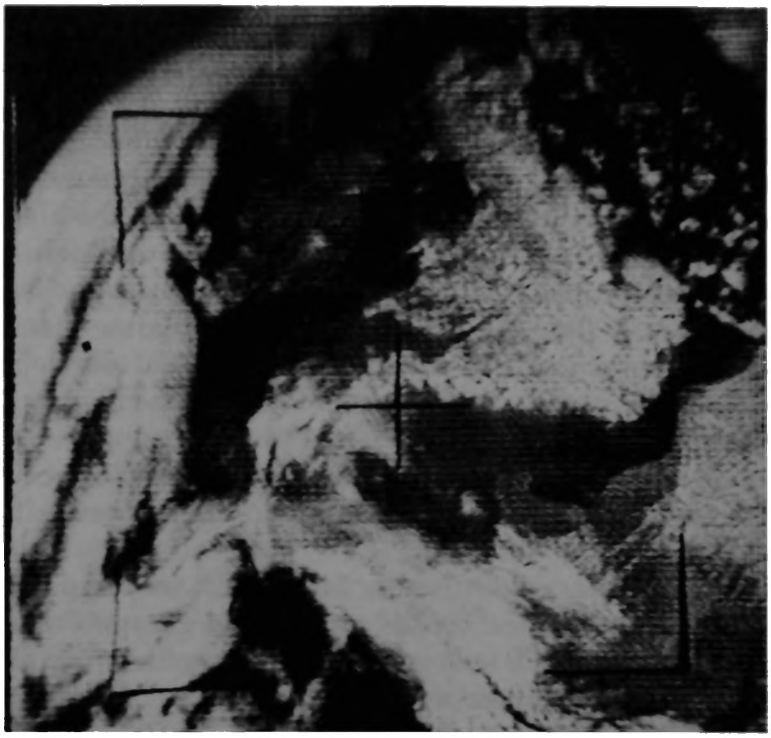
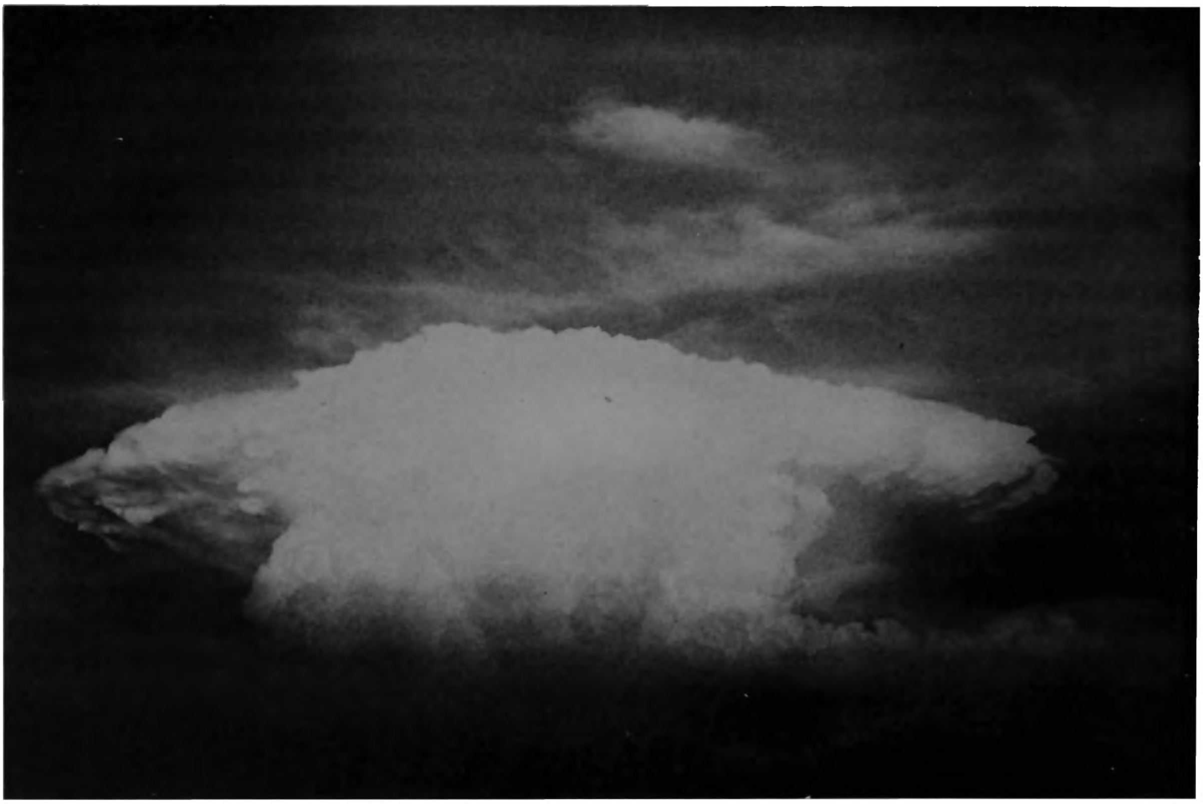


Figure 3.9
Afternoon cumulus over Great Britain as photographed by TIROS IV on April 13, 1962. The cloud cover to the south and west was associated with an approaching frontal storm. (NOAA, National Environmental Satellite Service)

Figure 3.10
Fully developed cumulonimbus (thunderstorm). The cloud extends through a vertical range of more than 6 km. (Official U.S. Navy Photo)



cumulus and *cumulonimbus*. Cumulus clouds are dense, dome-shaped, and have flat bases. They may grow to become cumulonimbus, the extent of vertical development depending upon the force of vertical currents below the clouds as well as upon the amount of latent heat of condensation liberated in the clouds as they form. Cumulus with little vertical development and a slightly flattened appearance are commonly associated with fair weather (Figures 3.8 and 3.9).

Sometimes large groups of cumulus develop a nearly complete covering of the sky and become stratocumulus. Conversely, stratocumulus occasionally break apart into separate clouds and become cumulus.

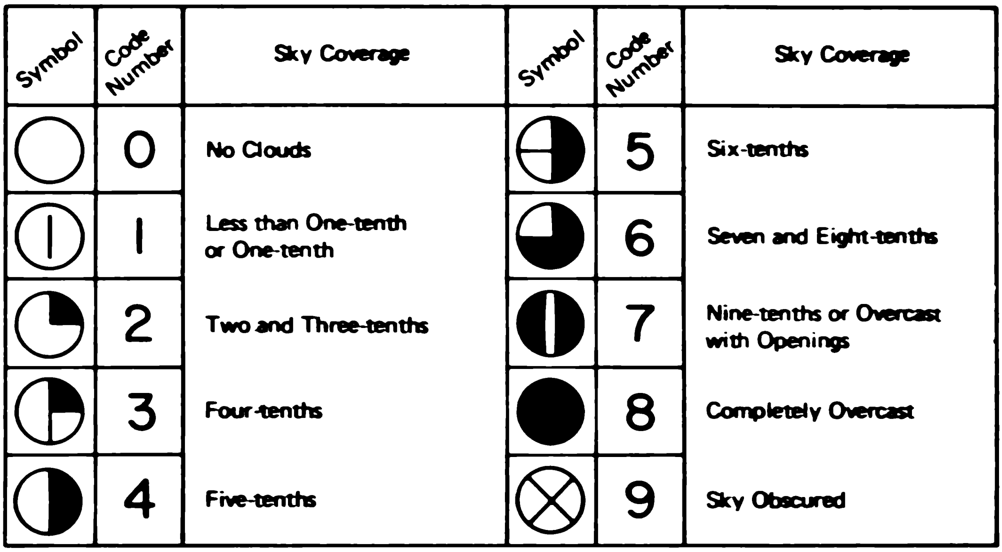
Cumulonimbus clouds exhibit great vertical development, towering at times to 18 km or more, where they spread out to leeward and form an anvil of cirrus (Figure 3.10). These are the thunderhead clouds that produce heavy showers of rain, snow, or hail, often accompanied by lightning and thunder. To an observer directly beneath, a cumulonimbus cloud may cover the whole sky and have the appearance of nimbostratus. In this case its true nature is revealed by the much heavier precipitation that falls from the cumulonimbus or by the preceding evolution of the cloud cover. Turbulence along the front of an advancing cumulonimbus sometimes creates a roll of dark fractocumulus or fractostratus.

CLOUD OBSERVATIONS

Because of their obvious relation to air masses, moisture content, and to vertical and horizontal movements of air, cloud observations are among the most valuable types of information used in weather forecasting and in flight planning. *Cloud types* are determined by visual observations, the accuracy of which depends upon experience and knowledge of the processes of cloud formation. The *sky cover* is obtained visually by viewing the clouds against the dome of the sky as a background and estimating the fraction of the sky covered. In the United States four terms are in common use to express sky coverage (see also Figure 3.11):

Clear, no clouds or less than $\frac{1}{10}$ of sky obscured by clouds
Scattered, $\frac{1}{10}$ to less than $\frac{6}{10}$ covered
Broken, $\frac{6}{10}$ to $\frac{9}{10}$ covered
Overcast, more than $\frac{9}{10}$ of sky covered

For meteorological records, the coverage of separate layers of clouds at different heights is determined and recorded. For example, an overcast of cirrostratus may be observed with $\frac{2}{10}$ of low clouds on the horizon. Of course it is not always possible to see upper cloud decks, especially



Symbol	Code Number	Sky Coverage	Symbol	Code Number	Sky Coverage
○	0	No Clouds	◑	5	Six-tenths
⦶	1	Less than One-tenth or One-tenth	◕	6	Seven and Eight-tenths
◔	2	Two and Three-tenths	◙	7	Nine-tenths or Overcast with Openings
◷	3	Four-tenths	●	8	Completely Overcast
◑	4	Five-tenths	⊗	9	Sky Obscured

Figure 3.11
Weather map symbols used to indicate sky cover on United States weather maps. (NOAA)

if they lie above a low overcast. Aircraft, satellites, and radiosonde observations greatly assist the ground observer in such cases.

Cloud height is the distance between the ground and the base of the cloud. The distance from the ground to the lowest cloud layer which creates a broken, overcast, or obscured sky cover is termed the *ceiling*. Cloud heights may be estimated visually, but for accurate determinations, one of several methods of measurement is employed. In the daytime, gas-filled *ceiling balloons* with a known rate of ascent are timed until they enter the cloud layer to determine height. Under a dark cloud layer or at night, the *ceiling light* is used to project a vertical beam of light onto the base of the cloud. The observer stands at a measured distance from the light projector and measures the angle of elevation of the light spot with a *clinometer*. Height of cloud is then determined by reference to trigonometric tables. The *ceilometer* operates on the same principle as the ceiling light but has a photoelectric element which reacts selectively to the light spot to indicate height (Figure 3.12). The accuracy of both the ceiling light and the ceilometer decreases with higher clouds and is lessened by a thin cloud cover since the spot of light becomes less definite.

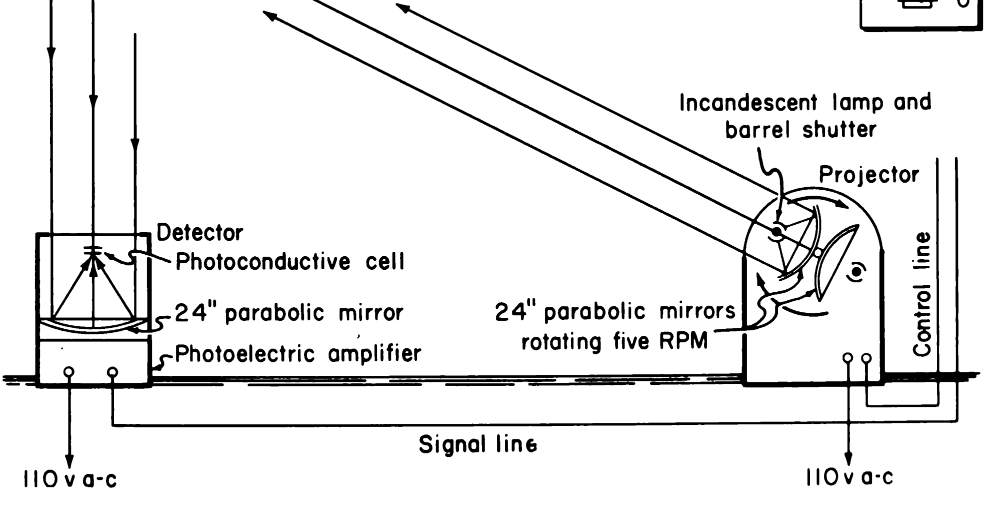
Cloud direction can be observed visually with reference to a fixed object, although for more accurate records the *nephoscope* is used. The principle of most nephoscopes involves a straight rod which is lined up with the direction of movement of clouds. To determine the *speed of cloud movement* the height of the cloud must be known. Movement with respect to graduations on the rod of the nephoscope is timed and by employing the mathematical principle of similar triangles cloud speed can be calculated. Some nephoscopes are fitted with mirrors so that clouds are observed indirectly.

Increasingly the composition and motions within clouds are being observed by means of radar techniques that can determine the size of particles and distinguish liquids from solids. Doppler radars measure droplet speeds and thereby assist understanding of cloud development. Lidar, an application of the laser, employs an optical principle to detect clouds and measure their height and rates of development. Observations from satellites provide extensive data on the global distribution of cloud cover, making it possible to relate clouds to the general atmospheric circulation and to storm systems.

FOG

Fog results when atmospheric water vapor condenses or sublimates (or when supercooled droplets freeze) to the extent that the new forms, water droplets or ice crystals, become visible and have their base in contact with the ground. Saturation of the air and sufficient nuclei are ordinarily

Figure 3.12
The rotating-beam ceilometer. (NOAA)



prerequisite to formation of fog, but, as has been previously explained, hygroscopic nuclei may "force" the process at relative humidities below 100 percent. Very light fogs formed in this way are sometimes referred to as *damp haze*. Heavier fogs restrict visibility and constitute a hazard. The collection of fog particles on vegetation and other obstacles may produce *fog drip,* which contributes to the receipt of moisture at the ground surface.

Observations of fog for meteorological records take into account the composition of the fog (water droplets or ice crystals) and its effect upon visibility. *Visibility* in a given direction is the greatest distance at which common objects like buildings or trees are visible to the unaided eye. *Prevailing visibility* is the greatest visibility value that prevails over at least one-half of the horizon. For determination of visibility at airport runways a device known as the *transmissometer* measures the transmission of light over a fixed path (Figure 3.13). Visibility can be affected by dust, smoke, and other suspended particles in addition to or apart from fog.

The principal processes that cause saturation are cooling of the air and evaporation of water into it. Accordingly, fogs are classified into two main categories:

1. Fogs resulting from evaporation
   (a) steam fog
   (b) frontal fog
2. Fogs resulting from cooling
   (a) radiation fog
   (b) advection fog
   (c) upslope fog
   (d) mixing fog
   (e) barometric fog

Steam fog forms when intense evaporation takes place into relatively cold air. As water vapor is added to the cool air saturation occurs and condensation produces fog. Steam fog is most commonly observed over bodies of water in middle and high latitudes. On the oceans it is also known as "sea smoke," and in the arctic builds to heights of 1,000 or 1,500 m. Steam fog sometimes forms on a small scale over warm, wet land immediately after a rain. In the tropics after a thundershower the air is temporarily cooled so that evaporation from soil and vegetation produces saturation and subsequent condensation.

Along the boundary between two air masses, evaporation from warm rain falling through the drier air below may be followed by saturation and condensation in cooler layers to form *frontal fog.* If the condensation layer is above ground level it is termed stratus cloud. (See

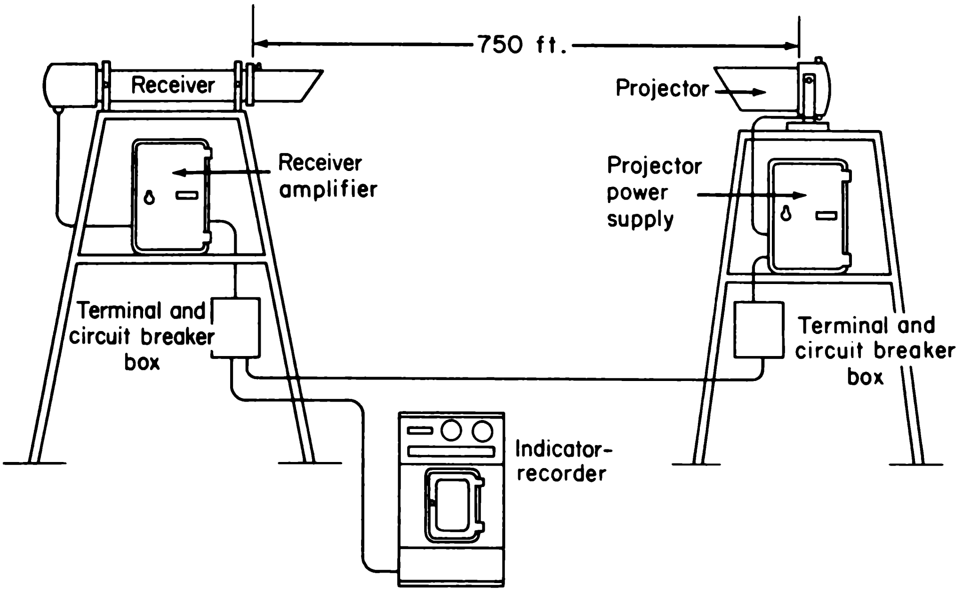


Figure 3.13
Schematic diagram of the transmissometer system used for determining airport runway visibility. (NOAA)

discussion of stratus as a low cloud.) Turbulence in the lower air reduces the extent and persistence of such a fog.

Ground fog, or *radiation fog,* is produced when fairly calm moist air is in contact with ground that has been cooled by nighttime radiation. If the air is completely calm only dew or frost is likely to form. Slight turbulence increases the depth of fog, but, if it is violent enough, it will distribute the excess moisture into upper, warmer layers and prevent fog formation, or dissipate that already in existence. Valleys are particularly susceptible to radiation fogs as well as to frosts because of frequent temperature inversions. At sea this type of fog does not form because of the negligible daily variation of the water temperature. In winter, largely in polar regions, and usually at temperatures below −30°C, radiation fogs may be composed of ice crystals. They are called *ice fogs* or sometimes *diamond dust* because the ice crystals glitter in the light.

Advection fog is formed when moist air is transported over a cold surface. It is especially common at sea where warm moist air masses move over the colder water of high latitudes or across cold ocean currents. In the summer, along arid west coasts, the cold water is provided by upwelling offshore; fogs of this type are frequent and persistent. Winds blowing onshore tend to carry along both temperature and humidity characteristics, extending the fog inland. In winter in mid-latitudes, the introduction of moist air of tropical or subtropical origin over cold land or snow surfaces likewise can result in advection fog.

The gradual orographic ascension of moist air up a sloping plain or hilly region can cool the air adiabatically to form *upslope fog*, provided the air is already near saturation. If the ascent is too rapid or there is convective turbulence in the air, condensation will likely take place above ground level to form clouds. Upslope fog is fairly common on the high plains east of the Rocky Mountains. In southeastern Wyoming it is locally known as "Cheyenne fog."

When warm moist air comes into contact with cool moist air the mixture in the boundary zone may have a temperature low enough to produce saturation and condensation. If this takes place at the earth's surface a *mixing fog* is the result. The most common occurrence is at fronts between air masses of maritime origin. Convection or convergence will cause it to dissipate or lift, and it is usually difficult to separate mixing as a cause from the other processes which generate fog at a front. When one "sees his breath" on a cold day, it is a special case of mixing fog, though it can hardly be regarded as a major fog type.

Barometric fog is extremely rare, although the cooling process associated with it possibly intensifies other types of fog. If the general pressure distribution over an area undergoes such a change that a layer of moist air at the ground level experiences a lowering of barometric pressure, the resultant adiabatic cooling could lead to condensation. These conditions are most probable in a valley or basin filled with stagnant air which is not immediately replaced as less dense air moves overhead.

The above outline of fog types should be regarded more as a classification of fog causes than as distinct kinds of fog. Although most fogs have a principal cause, their depth and density are usually influenced by one or more additional processes which lead to saturation. Figure 3.14 shows world areas having frequent fogs.

PRECIPITATION:
CAUSES, FORMS, PROCESSES, AND TYPES

Precipitation is defined as water in liquid or solid forms falling to the earth. It is always preceded by condensation or sublimation or a combination of the two, and is primarily associated with rising air. Although they contribute to the transfer of moisture from the atmosphere to the earth's surface, fog, dew, and frost are not considered to be precipitation. The common precipitation forms are *rain, drizzle, snow, hail,* and their modifications. Of these drizzle and light snow are the only ones likely to fall from clouds having little or no vertical development.

It is obvious that not all condensation, even in rising air, is fol-

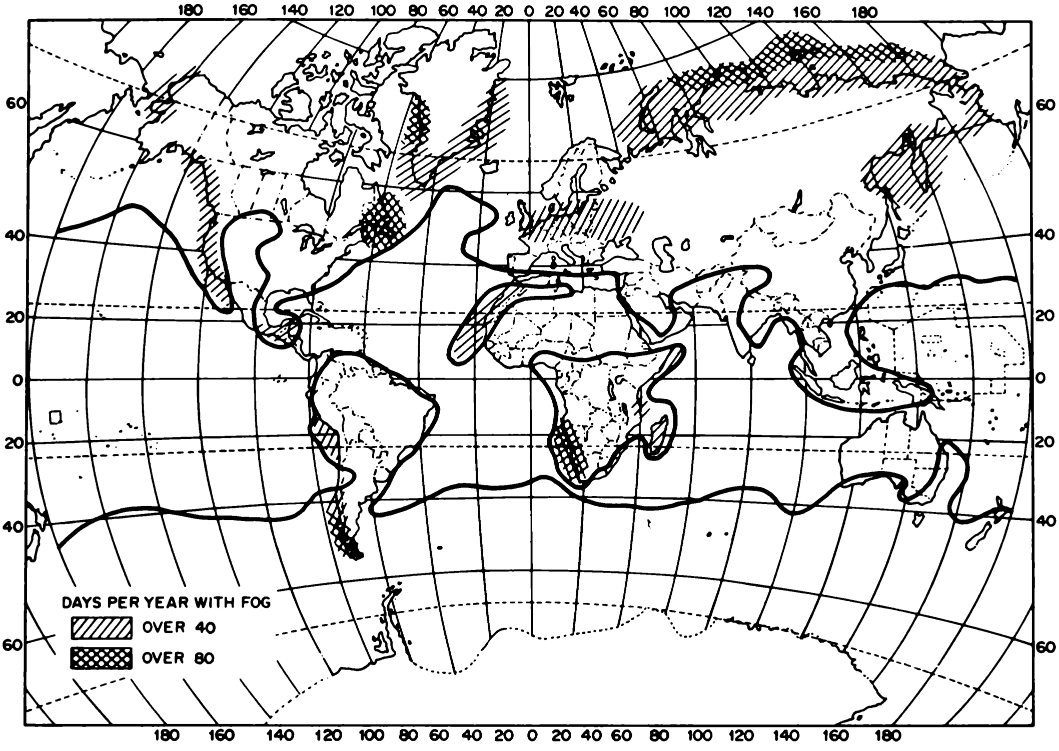


Figure 3.14
World areas having a high incidence of fog. The isolines represent a mean incidence of five days per year. (After Blüthgen; Van der Grinten projection, courtesy A. J. Nystrom & Co., Chicago)

lowed immediately by precipitation. A complete explanation of the mechanics by which some clouds produce rain or snow and others do not is still lacking. The "rainmaking" experiments carried out by various public and private agencies have considerably extended knowledge along these lines. A cloud is physically an *aerosol,* that is, a suspension of minute water droplets or ice crystals and other particles in air. In order to fall from the cloud, these water forms must grow to sizes which can no longer be buoyed up by the air. Coalescence may be achieved through collision resulting from turbulence or electrical discharge between droplets or, as is thought more likely, through the tendency of small droplets to migrate to larger, warmer to cooler, or liquid droplets to crystals. Until comparatively recently, the ice crystal theory of heavy precipitation was widely held, and it may be the explanation for many rainstorms in middle and high latitudes. In the upper levels of vertically developed clouds, there are often supercooled water droplets which do not play an active direct role in precipitation. If ice crystals appear in the supercooled cloud, water vapor crystallizes onto them and the water droplets evaporate into the resulting drier air. Ultimately all liquid water disappears and the crystals grow large enough to begin falling. If they fall through cold air, they may reach the ground as snow; if through warmer air, they melt to fall as rain. Unfortunately, this theory does not explain tropical precipitation which occurs from clouds having temperatures above freezing throughout. Nor can it explain the drizzle which falls from relatively thin stratus layers. Probably one or a combination of the other processes of coalescence is the basis of these types of precipitation.

Precipitation is classified in two ways: according to the form taken by the falling water or on the basis of the processes which lead to its formation. Rain is the most common precipitation form. It falls from clouds formed in rising air when the temperature, at least at lower levels, is above 0°C. As pointed out above, raindrops may begin as snow crystals which melt as they descend into warmer air. If the crystals grow and reach the ground level, they produce snow. Unusually large snowflakes are composed of several hexagonal crystals adhering to one another. Rain which freezes as it falls from a warm layer of air through a cold layer near the surface is termed *ice pellets,* formerly known as sleet. Another form of ice pellets consists of compacted snow covered with ice and falls in showers. Under wintertime conditions rain may fall through cold air and freeze upon striking cold surfaces to form *freezing rain.* Also known as glaze, ice storm, or "silver thaw," it ordinarily presages a warmer mass of air that moves in behind a warm front.

Drizzle consists of numerous uniformly minute droplets of water (less than 0.5 mm in diameter) that seem to float, following the slightest movement of the air. It falls continuously from low stratus-type clouds, never from convective clouds, and often accompanies fog and poor visibility. These characteristics distinguish it from light rain, which falls from convective clouds (generally nimbostratus) that have progressively lowered. Drizzle is called "Scotch mist," or simply mist in some places. When it strikes cold surfaces it may produce *freezing drizzle.*

Strong, rising convective currents, as in a cumulonimbus cloud, carry the raindrops formed by intense condensation and coalescence of water droplets to higher levels where they freeze to become *hail.* These frozen drops fall again after reaching a level of decreasing convection and take on a coating of ice as they pass through supercooled droplets. Repeated ascending and descending results in alternating layers of clear and crystalline ice and in a concentric structure in the hailstones. Another theory of hail formation holds that a frozen nucleus falls and alternately encounters supercooled water droplets and snow crystals which adhere to the hailstone and account for the concentric layers. The size of the stone depends on the amount of ice and snow it collects in one continuous descent. Hailstones having weights of more than a kilogram have been recorded, although they were probably two or more separate hailstones frozen together. An agglomerated hailstone that fell at Coffeyville, Kansas, on September 3, 1970, weighed 766 g (1.67 lb). Hail rarely falls in the tropics or at high latitudes.

Under less violent convection *snow pellets* may fall. They are small, opaque balls of compacted snow crystals, white in appearance, and do not have an ice coat. Snow pellets are more common in convective storms of winter and spring, whereas true hail comes with violent summer thundershowers.

Another ice form that may, in one sense, be regarded as precipitation is *rime*. Rime forms on cold surfaces that are exposed to moving supercooled fog or clouds, in contrast to frost, which is a sublimation product. It has a crystalline form and builds to the windward in featherlike shapes. A special case of rime is the rime icing of aircraft in supercooled clouds. It is most prevalent in temperatures below −9°C.

Having previously dealt with the methods of cooling which lead to condensation and sublimation, we can now outline the principal types of precipitation which result.

(1) *Convectional precipitation,* resulting from convective overturning of moist air. Heavy, showery precipitation is most likely to occur. Rain or snow showers, hail, and snow pellets are the forms associated with convective precipitation.

(2) *Orographic precipitation,* formed where air rises and cools because of a topographic barrier. It is doubtful that much of the world's precipitation is formed by orographic lifting alone; on the other hand, it is an important factor in intensifying rainfall on windward slopes and it therefore affects areal distribution. The greatest annual totals of rainfall in the world occur where mountain barriers lie across the paths of moisture-bearing winds. A famous example is Cherrapunji on the southern margin of the Khasi Hills in Assam, India. This station averages 1,144 cm of rainfall annually. In 1873 the total was only 719 cm; in 1861, when the annual total was a fantastic 2,299 cm, 930 cm fell in the month of July alone. Mt. Waialeale on the island of Kauai in the Hawaiian group has a windward-slope station that has an annual average of more than 1,200 cm. These and other record rainfalls in mountainous areas should not be ascribed entirely to a simple orographic effect. Besides forcing moist air aloft, orographic barriers hinder the passage of low pressure areas and fronts, promote convection due to differential heating along the slopes, and directly chill moist winds which come in contact with cold summits and snowfields. Even this combination of effects cannot induce precipitation unless wind and moisture conditions are favorable.

(3) *Frontal precipitation,* produced when air currents converge and rise. In tropical regions where opposing air currents have comparable

temperatures, the lifting is more or less vertical and is usually accompanied by convection. In middle latitudes, frontal convergence is characterized by the more gradual sloping ascent of warm air over cooler. However, convectional activity frequently occurs along fronts where the temperatures of the air masses concerned are quite different. Mixing of air along the front also probably contributes to condensation and therefore to the frontal precipitation.

Most precipitation results ultimately from a combination of cooling processes which produce condensation or sublimation. An additional process is necessary to cause raindrops or ice crystals to grow large enough to fall from the clouds.

OBSERVATIONS OF PRECIPITATION

A great deal of important information about precipitation is obtained by visual observation of its form, type, and duration. It is obviously impractical to measure all the precipitation which falls over large areas. Instead precipitation is sampled at a number of places. In the United States there are between 12,000 and 13,000 sampling stations. At a given place, precipitation is measured by the *rain gauge*. The standard rain gauge used in the United States is a metal cylinder 8 in. in diameter and has a capacity of 23 in. It is usually fitted with a covering funnel which directs the water into a measuring tube and which also reduces loss by evaporation. The measuring tube is so designed that it amplifies the amount of water at the ratio of ten to one. Thus 10 in. of water in the tube represent 1 in. of rain. A thin graduated stick is used to determine the depth to the nearest five-thousands of an inch, and amounts are recorded to the nearest hundredth inch. Alternatively, the water is poured into a graduated beaker. Amounts less than 0.005 in. (or 0.05 mm) are recorded as a trace, or simply "T." Values to the nearest tenth inch or whole millimeter are adequate for most climatological work. When snow, hail, or other ice forms are expected, the receiver funnel and measuring tube are removed from the gauge and only the 8-in. overflow can is used. In measuring snow and ice, a known quantity of warm water is added to the gauge. After melting is complete, the water is measured in the measuring tube and allowance is made for the warm water. Rain and snow tend to be blown in eddies about the gauge, making collection of a true sample questionable, especially in gusty winds. Slatted wind shields have been designed for mounting around the gauge to reduce this effect (Figure 3.15).

Improvements on the standard rain gauge are the *tipping-bucket gauge* and the *weighing-type gauge*. In the former a small divided metal

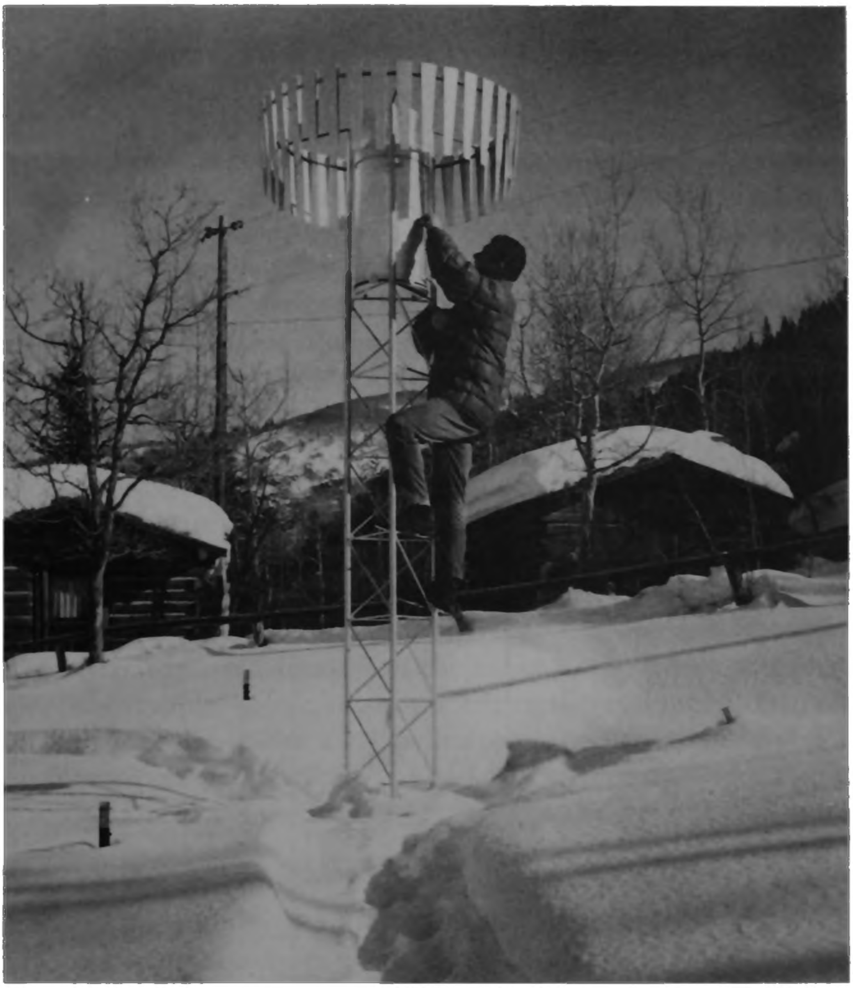


Figure 3.15
Precipitation gauge fitted with slatted shield to reduce effects of wind eddies.
(Bureau of Reclamation, U.S. Department of the Interior)

bucket is mounted so that it will automatically tip and empty measured quantities of rainfall. The tipping action activates electrical contacts which in turn activate a recording pen arm. The weighing-type gauge weighs precipitation as it falls and by means of a pen arm attached to the spring scale makes a continuous record on a clock chart.

Average snow depth on the ground is determined by averaging three or more representative measured depths. In mountain areas where snow depths become great, graduated rods are sometimes installed in representative spots so that depth can be read directly at the snow surface.

DISTRIBUTION OF PRECIPITATION REGIONALLY

Mean annual precipitation for the entire earth is about 86 cm (34 in.), but amounts vary greatly from place to place. In much the same way

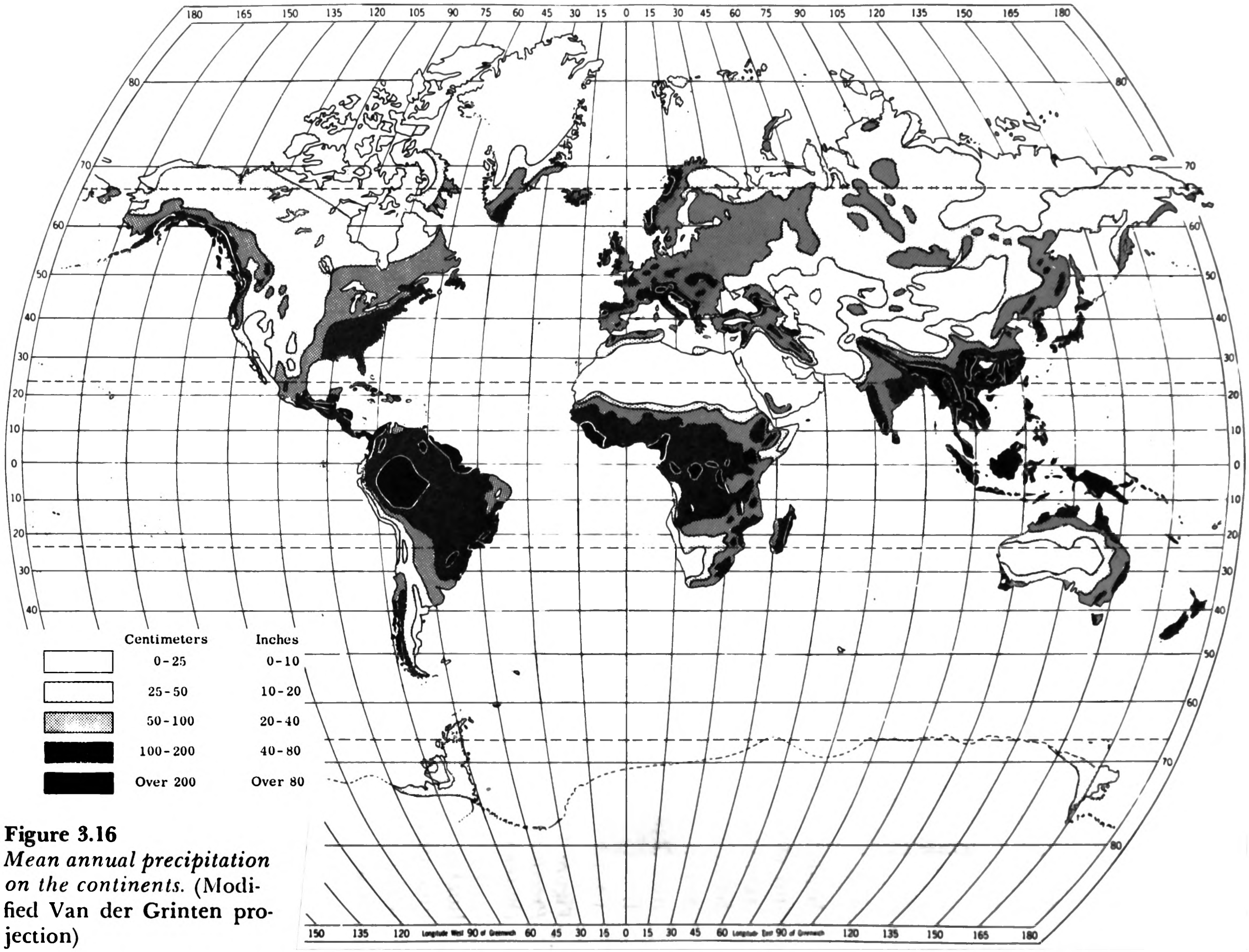


Figure 3.16
Mean annual precipitation on the continents. (Modified Van der Grinten projection)

that isotherms are used to show spatial distribution of temperature, values of precipitation may be plotted on a map using lines called *isohyets*. Figure 3.16 is an *isohyetal map* of the world with lines drawn through places with the same annual average rainfall. Note that precipitation over land only is indicated. Knowledge of rainfall over the oceans is extremely limited and for obvious reasons difficult to obtain. For land records, data for several years (at least 35) are desirable to provide representative mean values.

Although several climatic factors combine in a complex fashion to influence the world pattern, some generalizations are evident from the world map of precipitation. The amounts are greatest in equatorial latitudes, decreasing toward the poles with a great deal of irregularity. Where the general circulation of the atmosphere is characterized by ascent of air, for example, in the intertropical convergence zone, there are large annual rainfall totals. On the other hand, where subsidence is the rule, for example, in the subtropic highs, there is much less rainfall. In high latitudes the colder air has limited capacity for moisture, there are few incursions of moist tropical air, and precipitation is much less. There is also much less thermal convection in the polar regions than in the tropics.

The distribution of land and water imposes an important influence upon rainfall in middle and high latitudes. Regions located in the interior of continents, far from the oceanic sources of moisture, have less precipitation than coasts. The combination of nearby oceanic moisture sources with favorable prevailing winds produces greater rainfall on the windward than on leeward coasts, but ocean currents or drifts may introduce a further modification. Warm ocean currents offshore increase the moisture content of winds which blow across them. This principle operates along the Atlantic slope of the United States, where the warm Gulf Stream flows offshore. Cold currents, on the other hand, decrease precipitation over the adjacent land by cooling the lower layers of air that pass over them, thus inhibiting ascent and lowering moisture holding capacity. The result may be fog or low stratus clouds; vertical lifting and precipitation are unlikely.

Mountain barriers, by forcing ascent of moisture-bearing winds, tend to concentrate precipitation on their windward slopes and produce a "rain shadow" to the leeward. Aside from the tropics, the wettest regions are the mountain sides against which prevailing winds blow from the oceans. Even in the tropics, record annual rainfalls occur where orographic lifting is a significant factor. Among the numerous instances of mountains causing unequal distribution of precipitation are the mountainous trade-wind islands, the Atlas Mountains of North Africa, the Southern Andes along the Chilean-Argentine border, the Southern Alps of New Zealand, and the coastal ranges of western North America. Paradise Ranger Station in Mt. Rainier National Park, Washington, recorded 28.5 m (1,122.0 in.) of snow during the period July 1, 1971, to June 30, 1972.

The effect of orographic lifting on vertical distribution of precipitation also modifies its horizontal pattern. The increase at higher elevations is seen on the leeward as well as the windward slopes, but it appears to have an upper limit; precipitation begins to decrease again on the higher slopes. Much of the moisture has already been condensed and precipitated from the air at the higher elevations, and subsidence or downdrafts may induce evaporation of moisture-bearing clouds.

It should be noted that, because the precipitation of a given area usually results from several interrelated factors, its regional distribution is likewise determined by complex causes. "Normal precipitation" is largely a hypothetical concept with respect to both horizontal distribution and the time factor.

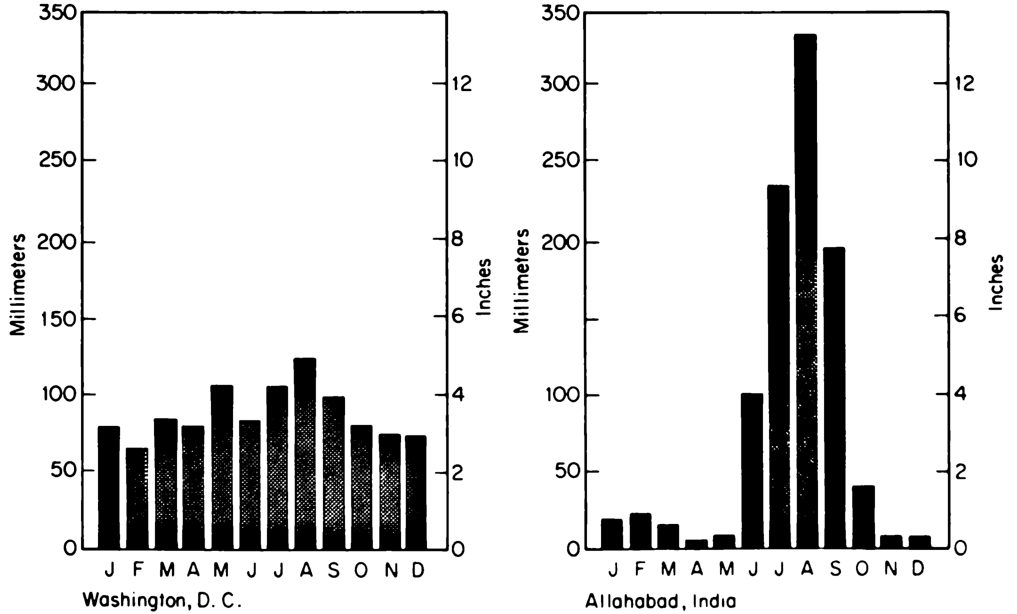
SEASONAL VARIATION OF RAINFALL

Quite as important as the actual rainfall totals accumulated in a particular year are the characteristics of *seasonal distribution* from month to month throughout the year, the *dependability* of rainfall from year to year, and the *frequency* of rainfall. The conditions which cause precipitation in a region do not exist in the same combination throughout the year. The latitudinal migration of the general pattern of wind and pressure belts is discussed in the section on general circulation in Chapter 4. Precipitation areas associated with the belts of convergence tend to move northward in the Northern Hemisphere in summer and southward in winter. In equatorial regions, which are constantly under the influence of equatorial convergence, rainfall is fairly heavy throughout the year, but a few degrees north or south are areas that are alternately dominated by the equatorial convergence and the subtropic high and have wet summers and dry winters. Certain west coasts in the middle latitudes have dry summers and wet winters as they experience the effects first of the subtropic high, and then of the westerlies with their cyclonic storms. The monsoon circulation brings even more striking seasonal contrasts, producing wet summers as winds blow onshore and dry winters when the circulation is reversed. Seasonal variation in rainfall because of the monsoon is especially well developed in India and southeast Asia. The effect of mountain barriers on seasonal variation can only be to increase the amount when the other factors are favorable. Because of its vital importance in human activities, especially agriculture, the annual *regime of precipitation* (monthly distribution through the year) is often more significant than the annual average. The precipitation graphs for Allahabad, India, and Washington, D.C., in Figure 3.17, clearly illustrate the difference in regimes although these stations have approximately the same annual averages.

Dependability of rainfall refers to the possible deviation from the average, that is, the variation from the normal. The actual annual total for a station may be far above or below the statistical annual average. There may also be a considerable deviation from the normal regime. Generally speaking, dependability of precipitation is relatively high in humid climates and decreases toward the regions with lower annual averages. The dependability of rainfall is a matter of great concern to farmers in subhumid and semiarid lands, where negative departure from the average annual rainfall—or even from the monthly means in certain critical months—can cause crop failure.

Frequency of rainfall is usually expressed as the number of precipitation days per year. Whether a station's total falls in a few heavy thundershowers or as light rain over a great number of days is of vital importance in many economic activities. Comparison of the amount of precipitation with the time during which it fell yields precipitation *in-*

Figure 3.17
Annual precipitation regimes at Washington and Allahabad.



Washington, D. C.
Annual average 1036 mm

Allahabad, India
Annual average 1032 mm

TABLE 3.2
World Maximum Observed Point Rainfalls*

Duration	*Millimeters*	*Inches*	*Location*	*Date*
1 min	31.2	1.23	Unionville, Maryland	July 4, 1956
5 min	63.0	2.48	Pòrto Bello, Panama	Nov. 29, 1911
8 min	126.0	4.96	Füssen, Germany	May 25, 1920
15 min	198.1	7.80	Plumb Point, Jamaica	May 12, 1916
20 min	205.7	8.10	Curtea de Arges, Romania	July 7, 1889
42 min	304.8	12.00	Holt, Missouri	June 22, 1947
9 hr	1,086.9	42.79	Belouve, Réunion	Feb. 28, 1964
12 hr	1,340.1	52.76	Belouve, Réunion	Feb. 28–29, 1964
24 hr	1,870.0	73.62	Cilaos, Réunion	Mar. 15–16, 1952
2 days	2,499.9	98.42	Cilaos, Réunion	Mar. 15–17, 1952
7 days	4,110.0	161.81	Cilaos, Réunion	Mar. 12–19, 1952
15 days	4,797.6	188.88	Cherrapunji, India	June 24–July 8, 1931
31 days	9,300.0	366.14	Cherrapunji, India	July 1861
2 mo	12,766.8	502.63	Cherrapunji, India	June–July 1861
6 mo	22,454.4	884.03	Cherrapunji, India	April–Sept. 1861
1 yr	26,461.2	1,041.78	Cherrapunji, India	Aug. 1860–July 1861
2 yr	40,768.3	1,605.05	Cherrapunji, India	1860–1861

**Daily Weather Map,* Dec. 13, 1962; J. L. H. Paulhus, "Indian Ocean and Taiwan Rainfalls Set New Records," *Monthly Weather Review,* 93, 5 (May 1965), 335.

tensity. This ratio may be expressed for an individual storm or for longer periods. Table 3.2 shows some observed rainfalls of exceptional intensity.

Additional Readings

DAY, JOHN A., "The Building Blocks of Clouds," *Natural Hist.,* 74, 1 (1965), 36–43.

ELLIOTT, W. P., R. EGAMI, AND G. ROSSKNECHT, "Rainfall at Sea," *Nature,* 229 (1971), 108–9.

International Cloud Atlas, 2 vols. Geneva: WMO, 1956.

LUDLUM, DAVID M., "Extremes of Snowfall in the United States," *Weatherwise,* 15, 6 (1962), 246–62.

MASON, B. J., *Clouds, Rain and Rainmaking.* London: Cambridge University Press, 1962.

———, *The Physics of Clouds,* 2nd ed. New York: Oxford Press, 1971.

NAGEL, J. F., "Fog Precipitation on Table Mountain," *Quart. Journ. Roy. Met. Soc.,* 82, 345 (1956), 452–60.

RUSKIN, ROBERT E., "The Measurement of Humidity in Meteorology," *Weatherwise,* 16, 2 (1963), 55–61.

SCORER, RICHARD, *Clouds of the World.* Harrisburg, Pa.: Stackpole Books, 1972.

SOBERMAN, ROBERT K., "Noctilucent Clouds," *Scient. Am.,* 208, 6 (1963), 50–59.

SUTCLIFFE, R. C., "Water Balance and the General Circulation of the Atmosphere," *Quart. Journ. Roy. Met. Soc.,* 82, 354 (1956), 385–95.

VAUGHAN, HARRY C., "The Spontaneous Freezing Temperatures of Melted Snow and of Small Water Drops," *Bull. Am. Met. Soc.,* 35, 2 (1954), 52–55.

chapter four

Motion in the Atmosphere

The atmosphere rotates with the earth, but it has complex internal motions of its own relative to the earth. Variable forces keep it in a state of constant agitation. One of the basic forces is air pressure, which is the main link between insolation and kinetic energy. Ordinarily we cannot perceive small changes in pressure, but together with other forces they motivate the winds that transfer heat and moisture from one area to another, often in connection with storms. Atmospheric motions also respond to the rotation of the earth, the effect varying with wind speed and latitude. Mountain barriers and surface friction influence both the direction and speed of air in motion, and when air streams come into conflict with one another their forces interact to complicate further the patterns of flow.

ATMOSPHERIC PRESSURE

Atmospheric pressure varies from time to time at a given place; it varies from place to place over short distances and on a worldwide scale; and it decreases with increasing altitude. The mass of a column of air above a point determines the atmospheric pressure at that point, but pressure also represents a force resulting from the kinetic energy of the gaseous molecules, and it is exerted in all directions. Horizontal pressure differences result primarily from temperature differences that produce density contrasts and from dynamic causes arising from atmospheric circulation. An increased proportion of water vapor in the air can alter pressure slightly because water vapor is less dense than the mixture of other constituents of air. Differences from place to place in the force of gravity also affect atmospheric pressure slightly.

Measurement of pressure variations is essential to the understanding of winds, storms, and related atmospheric phenomena. The most accurate instrument is the *mercurial barometer,* and most other pressure instruments are checked against it. The principle which operates the mercurial barometer is the balancing of the column of air against a column of mercury in a sealed glass tube (see Figures 4.1 and 4.2). Under standard conditions at sea level the atmosphere balances a column of mercury 76 centimeters (cm) or 29.92 in. high. Fluctuations in air pressure produce corresponding differences in the height of the mercury, and a graduated vernier is mounted along the tube to facilitate accurate reading to a thousandth of an inch. For determination of actual pressure values corrections are applied to adjust for the temperature responses of the mercury column and brass scale, errors in instrument construction, and the variation of gravitational force with latitude and altitude. The corrected reading is the *station pressure.* Obviously the station pressure on a high mountain will always be much lower than that at a nearby valley location, and a comparison of the two will be of questionable value in determining the lateral distribution of pressure. In order to make possible direct comparison of pressures at different altitudes station pressure is converted to *sea-level pressure.* It is assumed that the atmospheric column above the barometer extends on downward to sea level and a figure to represent the additional mass of air is added to the station pressure (or subtracted in those rare instances where the station is below sea level).

The height of the column in the mercurial barometer indicates the pressure in units of length, which are unsatisfactory for expression of a force. Measurements that employ the concept of pressure as a force are necessary for meteorological calculations. Accordingly, meteorologists use the *millibar* (mb), which represents the force exerted by 1,000 dynes on a square centimeter. (A *dyne* is the force which, acting on 1 g mass for 1 sec, imparts to it a velocity of 1 cm per sec.) In pressure measurements, 1 mb is equivalent to .0295299 in. of mercury. One inch of mercury represents the same atmospheric pressure at 33.86395 mb. The millibar value of mean sea-level pressure is 1013.2.

Physicists use *atmospheres* (atm) in laboratory work, 1 atm being the equivalent of 14.7 lb per in.^{2}, 29.92 in. of mercury, or 1013.2 mb. Table 4.1 shows some equivalents of the different pressure scales used in recording atmospheric pressure.

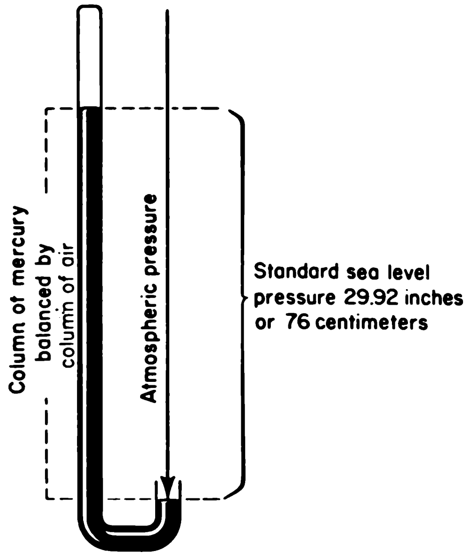

Column of mercury balanced by column of air
Atmospheric pressure
Standard sea level pressure 29.92 inches or 76 centimeters

Figure 4.1
Principle of the mercurial barometer.

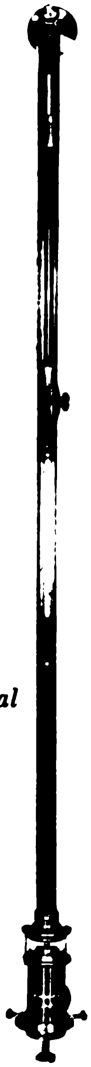


Figure 4.2
Fortin-type mercurial barometer. (NOAA)

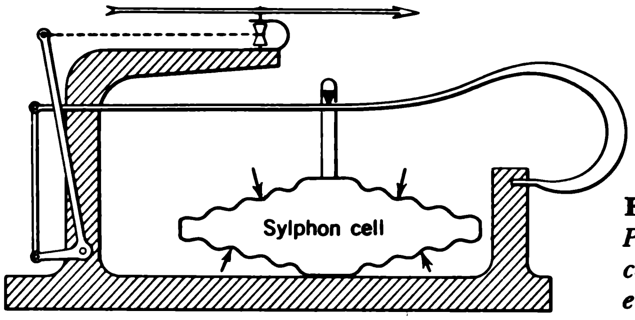


Figure 4.3
Principle of the sylphon cell in the aneroid barometer.

TABLE 4.1
Barometric Pressure Equivalents on Different Scales

Millibars	*Inches of Hg*	*Centimeters of Hg*	*Atmospheres*
940	27.76	70.51	.926
950	28.05	71.25	.937
960	28.35	72.01	.947
970	28.65	72.77	.957
980	28.94	73.51	.967
990	29.24	74.27	.977
1000	29.53	75.01	.987
1010	29.83	75.77	.997
1013.2	29.92	76.00	1.000
1020	30.12	76.50	1.006
1030	30.42	77.27	1.016
1040	30.71	78.00	1.026
1050	31.01	78.77	1.035

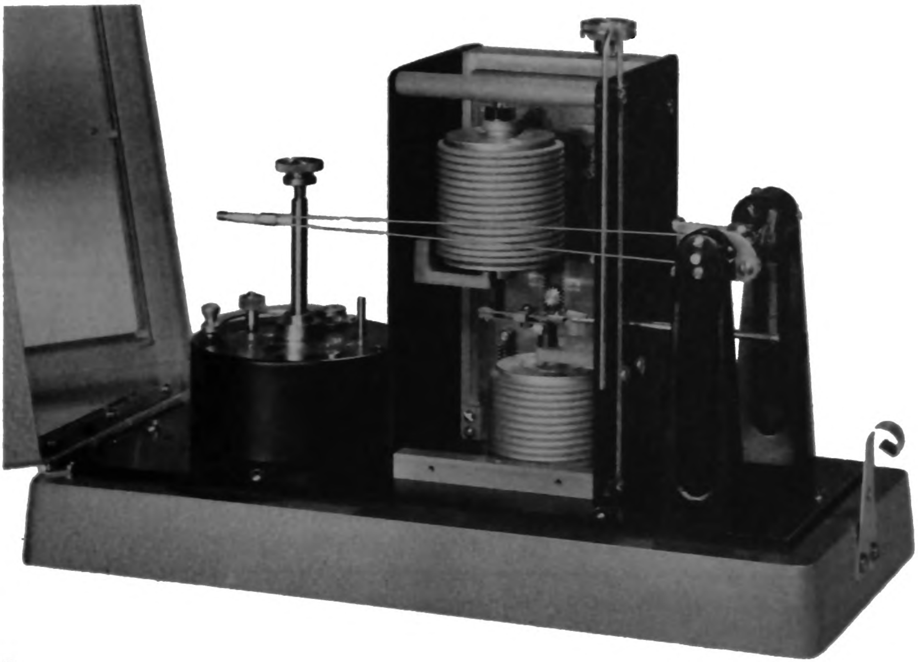


Figure 4.4
Microbarograph. Part of the assembly has been removed to show the sylphon cells. The pen arm traces a continuous record of pressure on a chart drum which sits on the clock at left. (Courtesy Belfort Instrument Company)

The *aneroid* barometer indicates pressure using the principle of the *sylphon cell,* which is a partially evacuated metal wafer (Figure 4.3). When pressure in the outside air increases, the cell tends to collapse; when pressure decreases, the cell expands. This fluctuation with pressure changes is mechanically linked to an indicator on a calibrated dial. The common aneroid barometer found in many households usually has a dial calibrated in inches of mercury and such terms as *rain, windy,* or *fair* printed along the circumference. Presumably the aneroid can forecast weather; actually pressure is only one of the weather elements, and it alone cannot be relied upon to indicate future weather with any certainty. The principle of the altitude barometer, or *altimeter,* is the same as that of the aneroid.

Another instrument employing sylphon cells is the *barograph.* It produces a continuous trace of barometric pressure on a clock-drum chart similar to that in the thermograph. Sylphon cells activate a pen arm which makes the ink record. Specially made *microbarographs* use several sylphon cells and a carefully constructed system of mechanical linkages to produce a more accurate record (Figure 4.4).

PRESSURE-HEIGHT RELATIONSHIPS

The density of air depends on its temperature, its composition, and gravity. Because all of these factors vary there is no simple relationship

TABLE 4.2
Standard Pressure-Height-Temperature Relations

Pressure (mb)	*Pressure (mm of Hg)*	*Altitude (m)*	*Temperature (°C)*
1013.25	760.0	Sea level	15.2
898.76	674.1	1,000	8.7
795.01	596.3	2,000	2.2
701.21	525.9	3,000	–4.3
616.60	462.5	4,000	-10.8
540.48	405.4	5,000	-17.3
265.00	198.8	10,000	–49.7
11.85	8.9	30,000	–41.8
0.26	0.19	60,000	-19.3

between altitude and pressure. Although atmospheric pressure always decreases with altitude, it does not always decrease at the same rate. Table 4.2 gives the average pressures and temperatures at selected levels under conditions that have been widely adopted as the *standard atmosphere* for the purposes of calibrating altimeters, for aerospace engineering, and for other scientific calculations. The standard atmosphere assumes a constant vertical structure of the atmosphere and is therefore an approximation of actual conditions.

HORIZONTAL PRESSURE DISTRIBUTION

Comparison of barometric pressure values that have been converted to a standard level reveals small horizontal differences that are highly significant in the analysis of weather patterns. On a worldwide scale mean sea-level pressure may vary from less than 990 mb to more than 1,030 mb. Agata, Siberia, recorded the world's greatest sea-level pressure on December 31, 1968, when the barometer reached 1,083.8 mb (32.005 in. of Hg). A record low of 877 mb (25.90 in. of Hg) was estimated from an aerial sounding in the eye of a typhoon west of the Mariana Islands on September 24, 1958.

To map the horizontal pattern of pressure we use *isobars*. An isobar is a line connecting points that have equal values of pressure. Maps of surface conditions ordinarily use sea-level pressure data, but they may also be drawn to show pressure distribution at a constant elevation in the upper atmosphere. Isobars are analogous to the contour lines on a relief map. In fact, another method of mapping pressure is the constant-pressure chart, on which contours indicate the heights of a given pressure value.

The underlying causes of most pressure differences at the bottom of the atmosphere are the same factors which affect the horizontal distribution of temperature, latitude and land-water relationships being the most important. It should be remembered that the terms *high* and *low* as applied to temperature and pressure are relative. Low pressures in the hot equatorial zone are not necessarily as low in actual value as the low pressures associated with middle-latitude cyclonic storms. The effect of latitude, through temperature, upon pressure is to produce a more or less symmetrical pattern of pressure zones on the earth. Along the equator lies a belt of low pressure known as the *equatorial low* or *doldrums*. In the cold polar latitudes are the vaguely persistent high-pressure areas, the *polar highs*. Centered at about 60° to 70° north and south latitude are the *subpolar low-pressure belts*, and at 25° to 35° north and south latitude are the *subtropic highs*. These intermediate pressure zones result ultimately from temperature differences, but it is not possible to regard temperature alone as the direct cause of pressure distribution, since wind plays an important part in redistributing density characteristics from one latitude to another. Furthermore, the global circulation includes vertical motions that may cause dynamic pressure differences. For example, the piling up of air by winds blowing along convergent paths, as happens at upper levels over the subtropic highs, increases surface pressure. Nor should these pressure "belts" be regarded as permanent. They are greatly affected by differences in net radiation resulting from seasonal migration of the sun and from variations in heating of land and water surfaces. All are subject to incursions of air masses with contrasting temperature and density properties from other latitudes.

The effect of the irregular distribution of land and water upon pressure is best seen in seasonal contrasts, especially in the middle latitudes. In winter the continents are relatively cool and tend to develop high-pressure centers; in summer they are warmer than the oceans and tend to be dominated by low pressure. Conversely, the oceans are associated with low pressure in winter and high pressure in summer. By reference to Figures 4.5 and 4.6, which show the average sea-level pressures for January and July, it can be seen that the North Atlantic, for example, is under the dominance of high pressure in summer and low in winter. In contrast, the Southern Hemisphere, which has a more homogeneous surface (mostly water) does not show such great seasonal differences. Note that the isobars in the latitudes 40° to 70° south are nearly parallel and aligned along parallels of latitude. The tendency to greater persistence of the subtropic high and the more orderly arrangement of isobars in the Southern Hemisphere are both results of the smaller effect of land-water differences.

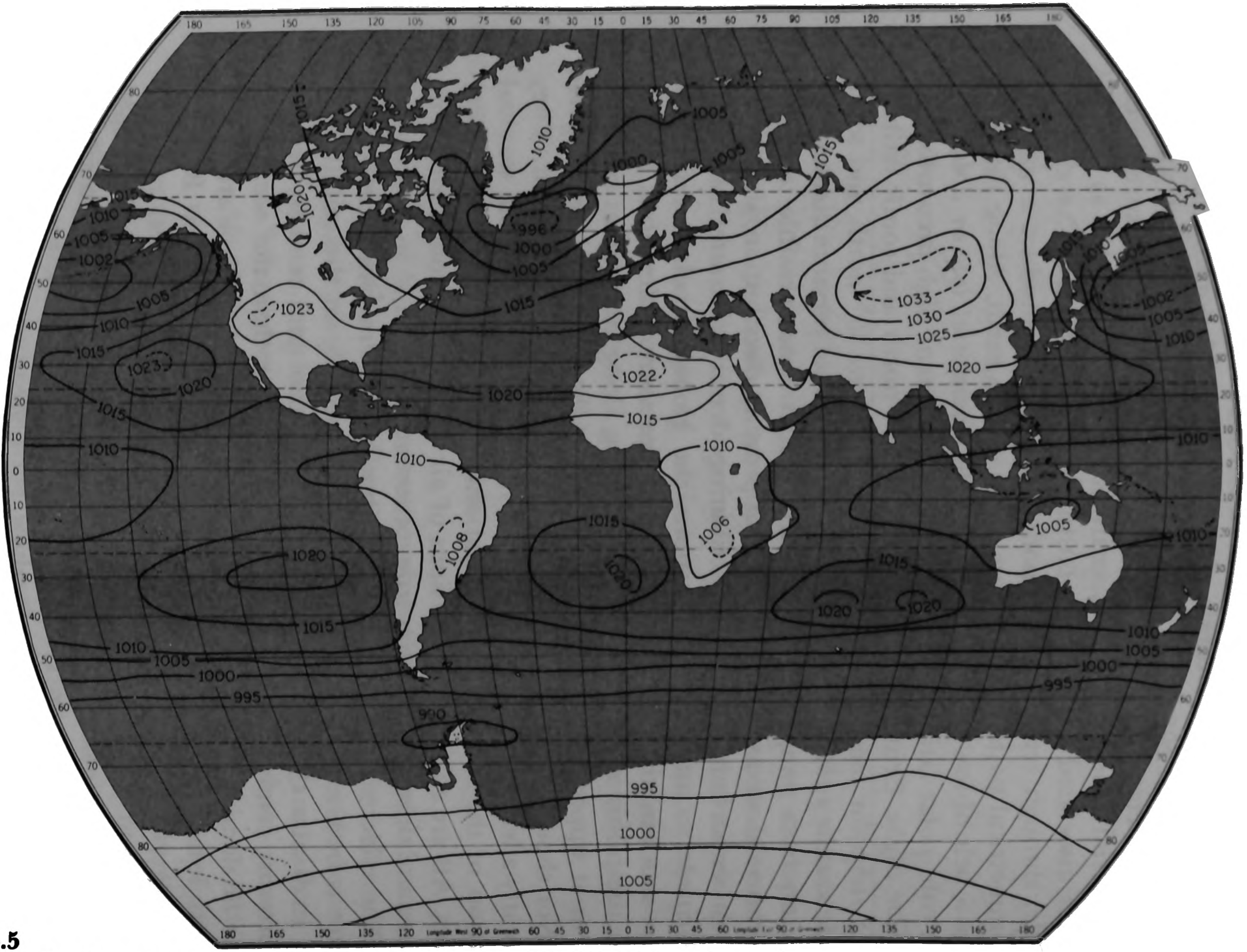


Figure 4.5
World distribution of sea-level pressure in January. (Modified Van der Grinten projection)

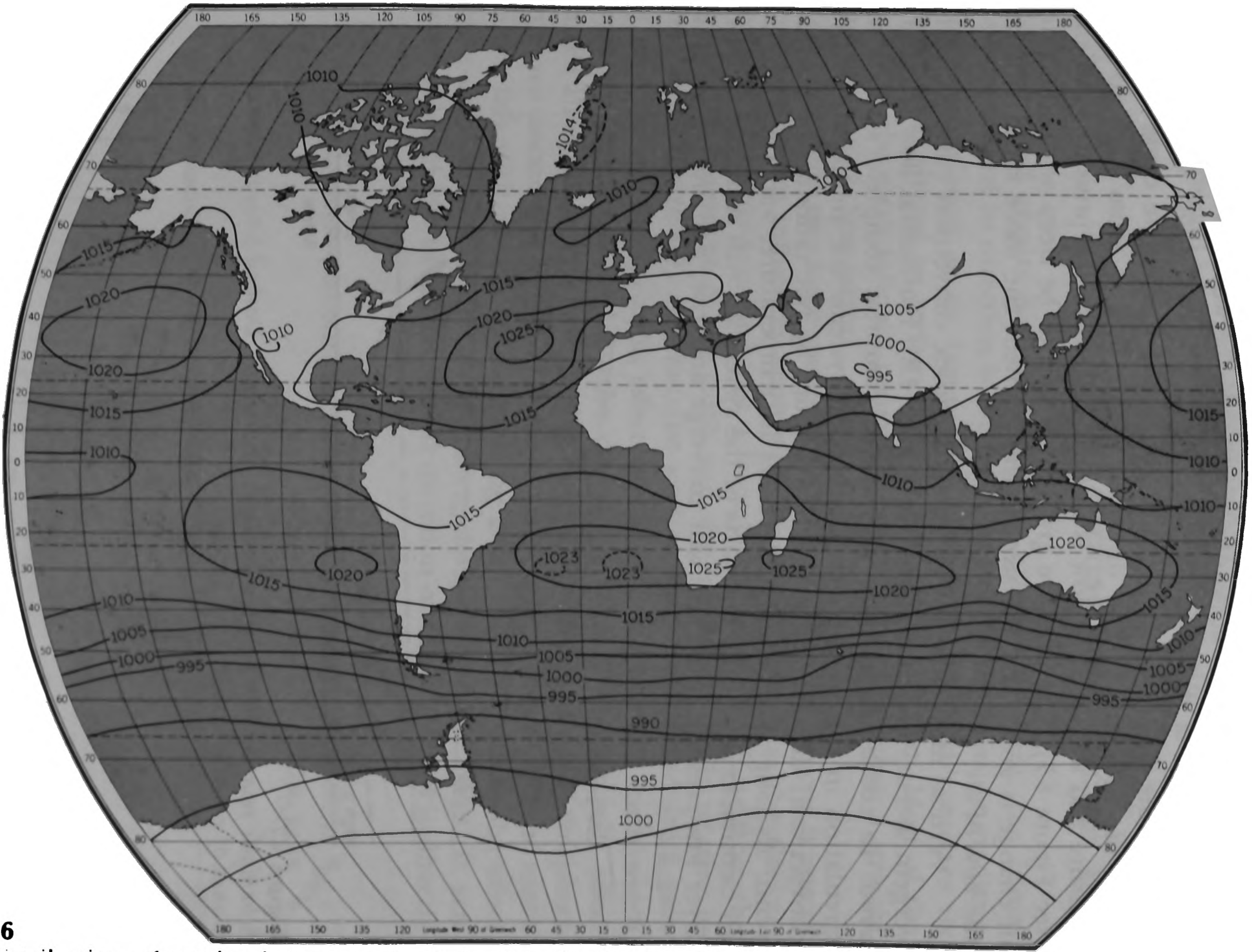


Figure 4.6
World distribution of sea-level pressure in July. (Modified Van der Grinten projection)

PRESSURE AND WIND

The rate of change in atmospheric pressure between two points at the same elevation is the *pressure gradient* or *isobaric slope*. It is proportional to the difference in pressure and is the immediate cause of horizontal air movement. The direction of air flow is from high to low pressure and the speed of flow is directly related to the pressure gradient (see Figure 4.7). The pressure gradient is said to be steep when the rate of change is great, and the steeper the gradient the more rapid will be the flow of air.

Vertical movements of air include eddies, convection currents, convergent ascent, and subsidence (see Figure 4.8). Horizontal flow is called *wind*. In combination these movements comprise complete systems of circulation in the atmosphere. Convection systems are analogous to the convectional distribution of heat in a room by hot radiators or hot air vents. As long as sufficient differences in temperature, and therefore of air density, exist at different parts of the system air will continue to flow. Over low-pressure areas, air will converge and rise, and over high-pres-

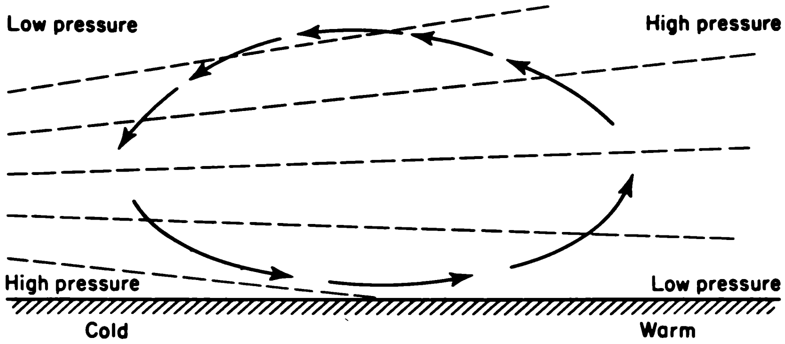


Figure 4.7
Schematic relationship between pressure and winds. Dashed lines represent planes of equal pressure.

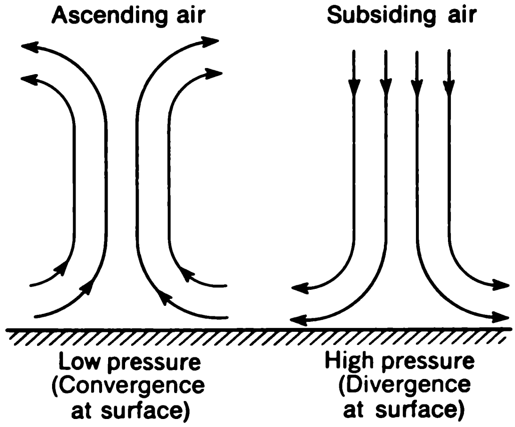

Ascending air
Subsiding air
Low pressure (Convergence at surface)
High pressure (Divergence at surface)

Figure 4.8
Schematic relationship between vertical air currents and pressure.

sure areas, the air will subside and diverge. Strong gradients between the two are indicated by closely spaced isobars on a pressure map, and the resulting winds will be of correspondingly high velocity.

FACTORS AFFECTING WIND DIRECTION AND SPEED

Because wind results basically from the pressure gradient, the initial determinant of its direction is the force exerted by the pressure gradient. However, as soon as air begins to move along the earth's surface its direction is altered by certain other factors acting together. Most of the winds of the earth follow a generally curved path rather than a straight one because of these factors.

The deflective force due to the earth's rotation on its axis is one of the most potent influences upon wind direction. Known as the *Coriolis force,* it is, strictly speaking, not a force but an effect resulting from the rotational movement of the earth and the movement of air relative to the earth (Figure 4.9). But because we live on the earth and are a part of its rotation, the apparent effect is that of a force which turns winds from the paths initiated by the pressure gradient. The Coriolis effect causes all winds (indeed all moving objects) in the Northern Hemisphere to move toward the right and those of the Southern Hemisphere to move to the left with respect to the rotating earth. At the equator the effect has a value of zero and it increases regularly toward the poles. It acts at an angle of 90° to the horizontal direction of the wind and is directly

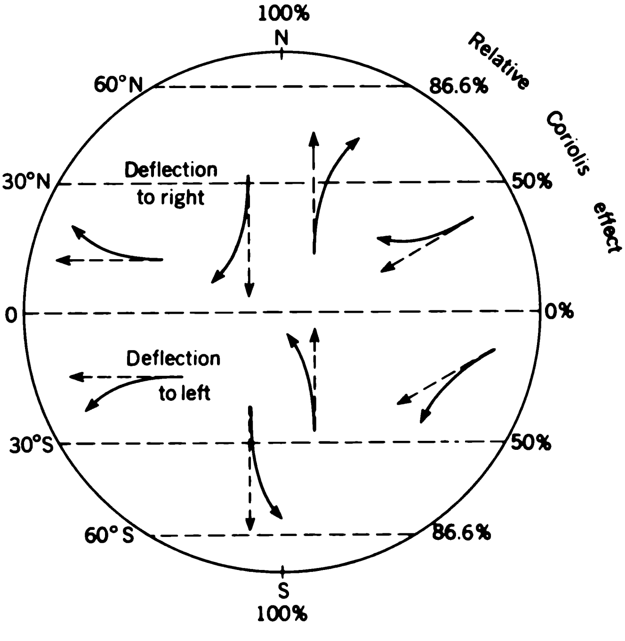


Figure 4.9
Deflection of winds by the Coriolis effect.

proportional to horizontal wind speed. When the pressure gradient has initiated movement of air the resulting wind is deflected more and more to the right (left in the Southern Hemisphere) until it may be blowing parallel to the isobars, that is, at right angles to the pressure gradient. If the motion of the air is along curved isobars a net centripetal acceleration tends to pull it toward the center of curvature, producing the *gradient wind* and a rotating motion relative to the earth's surface. A counterclockwise flow in the Northern Hemisphere is termed cyclonic; clockwise flow is anticyclonic. (See Figure 4.10).

Along and near the earth's surface wind does not move freely in a horizontal plane. Irregularities in surface relief and local differences in thermal convection cause moving air to take on correspondingly irregular motion so that it undergoes abrupt changes in speed and direction. This fluctuating wind action, known as *turbulence,* is associated with lulls, gusts, and eddies and increases with increasing wind speeds. The effects of surface turbulence are not very great above 500 m, and at about 1,000 m actual wind direction and speed are equivalent to the theoretical gradient wind, which may be calculated from the pressure gradient and other forces. The effect of surface friction is to reduce wind speed, in turn reducing the Coriolis effect so that the wind moves across the isobars at an angle toward lower pressure.

Other factors being equal, the difference in wind speed and direction between the surface and upper levels is greatest over rough land surfaces. Over water the surface wind more nearly approximates the gradient wind. On the average, low-speed winds cross the isobars over land at an angle of about 45° with a speed about 40 percent of the gra-

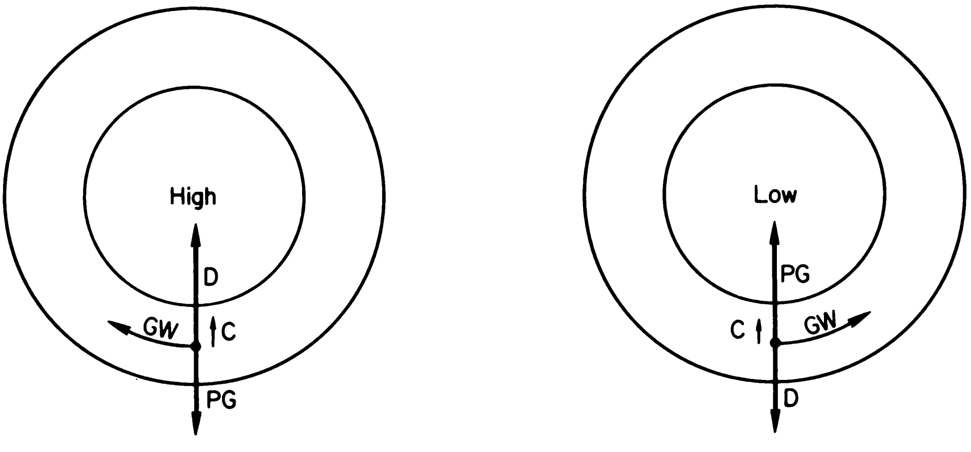


Figure 4.10
Forces affecting the gradient wind around high- and low-pressure centers. PG, *pressure gradient;* D, *deflective (Coriolis) effect;* C, *centripetal acceleration;* GW, *gradient wind resulting from balanced forces.*

dient wind speed; over oceans the angle is about 30° and the surface speed about 65 percent of the gradient wind speed. Recognizing the relationship of wind direction to pressure distribution, Buys-Ballot formulated the rule: If you stand with your back to the wind in the Northern Hemisphere, pressure is lower on your left than on your right. In the Southern Hemisphere, again with your back to the wind, lower pressure will be on your right and higher pressure on your left.

WINDS ALOFT

In free air the effect of increasing altitude is to reduce the kind of turbulence induced by surface friction. Ordinarily wind direction changes with altitude so as to become more nearly parallel to the isobars and wind speed more nearly reflects the pressure gradient. Strong convection currents may cause turbulence at high levels, however. Along boundaries between air streams having converging directions or different speeds *wind shear* develops, sometimes creating violent turbulence. The swift passage of wind over the top of relatively calm air in a temperature inversion is likely to produce a turbulent layer.

Considerable variation of wind direction with altitude may result from convergence of two air masses, particularly if their temperatures are different. Along the boundary where cold air advances against warmer air the lighter, warmer air will rise and tend to spread out above the heavier, cold air so that winds in the upper air may be blowing directly opposite the cold surface winds. This phenomenon is commonly associated with the cyclonic storms of middle latitudes but it is by no means confined to those areas.

Because local winds are all comparatively shallow and have pressure gradients only vaguely related to the general circulation, vertical soundings through and above them often show great variation of both direction and speed. If the conditions responsible for the local wind create a pressure gradient in the same direction as the major circulation, the two will simply reinforce one another.

Although the considerations of upper winds so far have been chiefly with respect to free air, surface winds on hills and mountains are governed by most of the same principles. Rising above the maximum friction layer, mountain peaks are buffeted by the higher-speed winds of upper levels. The condition of trees near the timberline and the experiences of mountain climbers attest to this. The summits of mountains tend to increase further the speeds at upper levels. Air blowing across the top of a mountain is constricted between the summit below and the layers of air above; being thus funnelled, it increases in speed until it passes the summit. A dramatic example of wind force on a mountaintop occurred at Mt. Washington in northern New Hampshire on April 12, 1934. Wind speeds of more than 130 knots (67 m per sec) were recorded for most of the afternoon and at one time reached 200 knots (103 m per sec).

DIURNAL VARIATION OF WIND SPEED

If we consider the air movement in an extensive wind system and disregard wind shifts or locally induced winds, we find a fairly regular daily variation of wind speed at the surface. The maximum speed usually occurs in the early afternoon and the minimum in the early morning hours just before sunrise. Probably everyone has experienced the phenomenon of winds dying down in the evening only to rise again with renewed force in the morning. In order to find the reason for this it will be necessary to recall that wind speed increases with increasing altitude. During the daytime, convection caused by heating of the surface air layers brings about an exchange of air between lower and higher levels, and a more nearly uniform vertical distribution of speed exists. At night the air near the ground is cooled and, being heavy, will tend to remain at the low levels, where because of the greater frictional drag it will resist being carried along by the fast-moving winds above. On high hills and mountains, especially if they are isolated, the effect may be the reverse. During the day the slowing effect of frictional drag is transferred to the upper levels by turbulence to produce a midday minimum of wind speed. At night the air rides above the cool layer lying in depressions and reaches its maximum speed for the 24-hour period. This is illustrated by the graphs in Figure 4.11, which compare wind speeds at San Francisco with those of nearby Mount Tamalpais. Note that in July, when daytime heating is much greater, the diurnal variation of wind speeds is greater. In the case of San Francisco fog often retards surface heating so that a westerly sea breeze does not reach its maximum speed until late afternoon.

WIND OBSERVATIONS

Wind direction may be ascertained quite simply by watching the movement of clouds, vegetation, smoke, waves on water surfaces, and so forth. This type of visual observation yields important information on winds, and, with some experience, an observer can also estimate the speed of surface winds. In 1805 Admiral Francis Beaufort introduced a wind

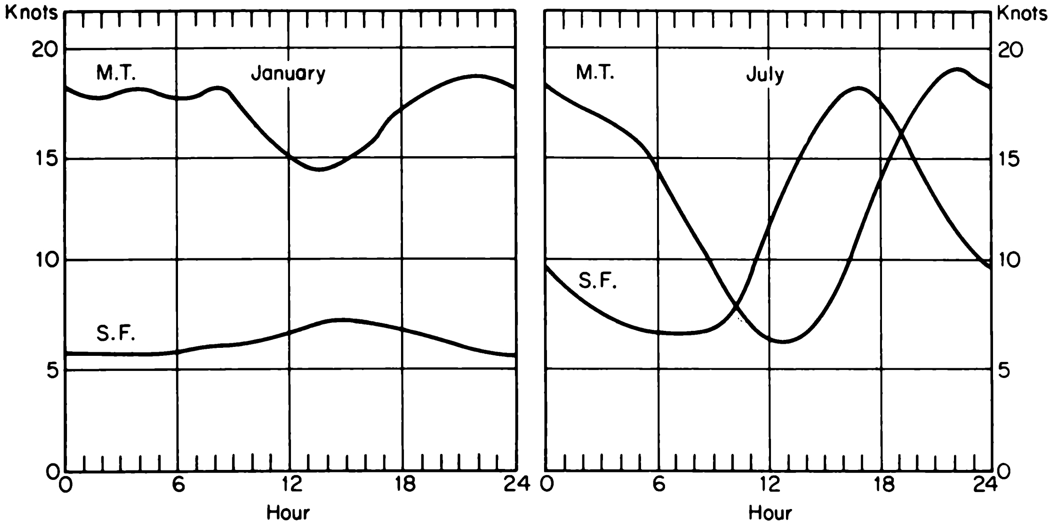


Figure 4.11
Hourly variations in mean wind speeds at Mount Tamalpais (794 m) and San Francisco (67 m) in January and July. (Data from Herbert H. Wright, "Characteristics of Winds at Mount Tamalpais, Cal.," *Mon. Wea. Rev.,* 44, 9 (1916), 512–14)

force scale which was based upon the response of certain objects to the wind. In applying the Beaufort Scale the extent to which smoke is carried horizontally or to which trees bend before the wind is used as an index of speed. At sea the condition of waves, swell, and spray in addition to response of sails and masts is the basis for windspeed estimates.

Most weather stations are equipped with a wind observation unit which consists of a wind vane and an anemometer mounted at the top of a steel support column. The wind vane points into the wind, but it should be noted that winds are named for the direction from which they come. A wind blowing from north to south is a north wind. Meteorologists commonly use degrees of azimuth measured from north a +0°, around through east (90°), south (180°), and west (270°). North is represented by both 0° and 360°. Ordinarily the wind vane is electrically connected to show wind direction on an indicator inside the weather station, since the instrument support may have to be mounted at some distance from the roof immediately above the station. Adaptation of the movement resulting from changing wind direction to a pen arm makes possible a continuing record on the *wind register.*

In the United States and certain other countries the standard unit of wind velocity for observations is the *knot,* that is, 1 nautical mi per hr. One knot equals 6080.20 ft, or about 1.15 mi, per hr; 1 mi per hr equals about 0.87 knots. The use of smaller units such as feet per second or meters per second is ordinarily restricted to research or microclimatic observations. One knot equals 0.5148 m per sec.

Anemometers are instruments for the measurement of wind speed (see Figure 4.12). The most common type is the Robinson cup anemometer in general use by the United States National Weather Service. Either three or four metal hemispheres are mounted on arms which rotate freely about a vertical axis. A system of gears similar to an automobile speedometer translates the rotation generated by the wind to an indicator or recorder for both speed and miles of wind that have passed the instrument.

The *bridled anemometer* consists of a series of cups (usually 32) mounted somewhat as in the cup anemometer, but the vertical shaft is held, that is, bridled, by a spring. The rotation of the shaft against the spring represents wind force and is translated to an indicator which shows wind speed. This instrument is commonly used on ships. In order to determine wind velocity, however, allowance has to be made for the direction and speed of the ship.

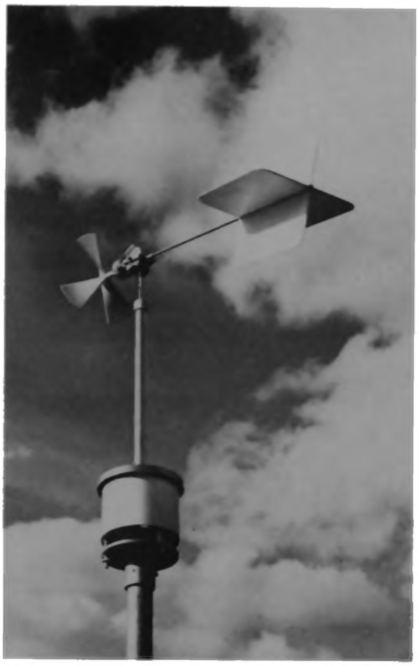
The *pressure-tube anemometer* has the same principle as the *pitot tube* used on airplanes to determine air speed. It is mounted as a wind vane to keep the open end of the tube pointed into the wind. The difference between pressure due to the force of the wind and the atmospheric pressure activates a pressure gauge calibrated for wind speed. The dynamic principle employed in this instrument causes it to lag in registering gusts.

Air movement can also be measured electrically. The rate of cooling of an exposed hot wire, a thermistor, or a thermocouple is a function of air movement and therefore an index of wind speed. Vertical as well as horizontal speed and direction of air flow is important in microclimatological studies. Bivanes and various assemblies that are sensitive to forces from all directions provide the desired information.

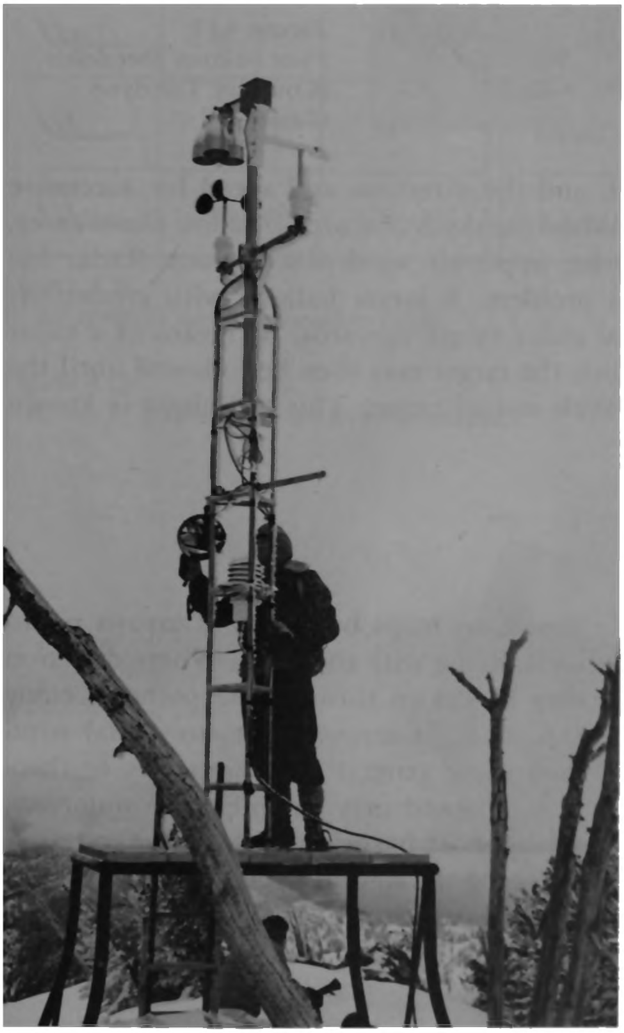
For purposes of analyzing storms and air masses as well as for planning air travel, observations of upper winds are as important as those of winds at the surface. Clouds are an obvious indicator of direction of winds aloft, and if their height is known their drift can be used to estimate their velocity.

The most common method of obtaining upper wind data is by means of the *pilot balloon,* or "pibal." A balloon filled with hydrogen or helium is released from the earth's surface and as it floats aloft it rides the winds to reveal their direction and speed. The *theodolite,* a right-angled telescopic transit which is mounted so as to make possible reading of both azimuthal (horizontal) and vertical angles, is used to track the balloon (Figure 4.13). The latter's progress is plotted minute by minute

Figure 4.12 *Instruments for wind measurements.* (A and C courtesy R. M. Young Company; B courtesy Bureau of Reclamation, U.S. Department of the Interior)

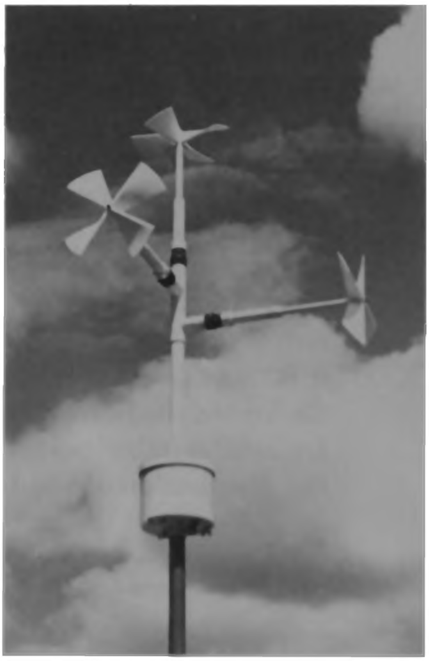


A, *anemometer bivane. The vane fins move the propeller to measure vertical as well as horizontal wind components.*



B, *cup anemometer and wind vane. Infrared lamps prevent freezing of the anemometer.*

C, *UVW anemometer. Propellers mounted at right angles respond to air movement in three dimensions.*



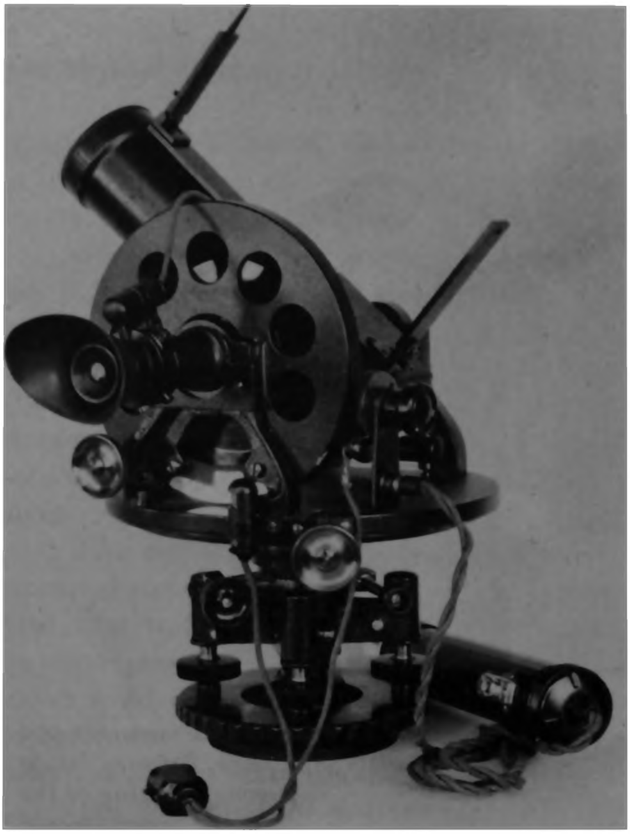


Figure 4.13
Pilot balloon theodolite.
(Courtesy Teledyne Gurley)

on a special plotting board, and the direction and speed for successive altitudes can be computed. When the sky is obscured by a low cloud cover, the pilot balloon is useless for upper-air wind observations. Radar has provided the answer to this problem. A larger balloon with greater lift is employed to carry a metal radar target upwards. By means of a radar transceiver or radio theodolite, the target may then be followed until the balloon either breaks or travels out of range. This technique is known as a *rawin* observation.

MAPPING WIND DATA

Wind direction is ordinarily shown on maps by means of arrows pointing in the direction of flow, that is, flying with the wind. Where direction only is indicated, the arrow may be drawn through the point or circle representing the station location. Straight arrows represent actual wind observations; curved *stream lines* show general air trajectories or theoretical flow patterns. Although wind speed may be entered in numerals, it is more commonly designated by short barbs or feathers at the tail of the arrow, the head of the arrow being the station circle (see Figure 4.14). *Isotachs* connect points that have equal wind speed values and show the distribution of wind force on maps.

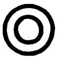
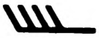
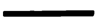
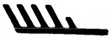
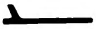
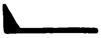
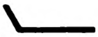
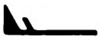
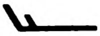
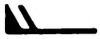
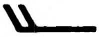
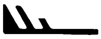
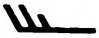
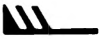
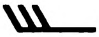
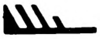
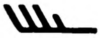
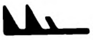
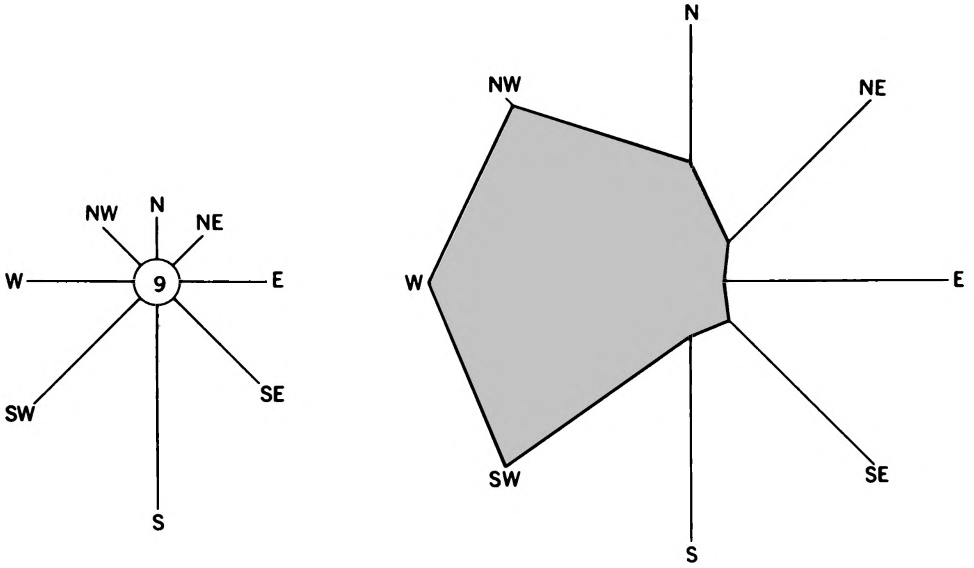
Symbol	Miles (Statute) Per Hour	Knots	Symbol	Miles (Statute) Per Hour	Knots
	Calm	Calm		44 - 49	38 - 42
	1 - 4	1 - 2		50 - 54	43 - 47
	5 - 8	3 - 7		55 - 60	48 - 52
	9 - 14	8 - 12		61 - 66	53 - 57
	15 - 20	13 - 17		67 - 71	58 - 62
	21 - 25	18 - 22		72 - 77	63 - 67
	26 - 31	23 - 27		78 - 83	68 - 72
	32 - 37	28 - 32		84 - 89	73 - 77
	38 - 43	33 - 37		119 - 123	103 - 107

Figure 4.14
Wind symbols used on weather maps.

Figure 4.15
Methods of graphing mean wind directions.



Wind rose Valentia, Ireland (January)
(Number in circle represents
percentage of calms)

Wind direction frequency polygon
Boston, Massachusetts (January)

Wind roses portray climatic aspects of winds graphically on maps, charts, and diagrams. The relative lengths of lines radiating from the station circle indicate the percentage frequency of principal wind directions (see Figure 4.15). The same data may be plotted as a polar graph to display a *wind frequency polygon.*

LOCAL WINDS

In order to develop an understanding of the factors that influence wind systems it is helpful to begin with relatively simple small-scale motions. A number of winds of local importance develop from local temperature differences. They usually affect small areas and are confined to the lower levels of the troposphere. Their air movements may be generated either by heating or cooling of a particular area. Sometimes they represent virtually complete convectional circulation; in other cases they merge into the stronger wind systems of the general circulation or of major storms.

The *sea breeze* is one of the common local winds. Along seacoasts or large inland water bodies in summer the land heats much faster than the water on a clear day and a pressure gradient is established, directed from high over the water to the low over the land (Figure 4.16). The circulation which follows brings cool air onto the land, but the system is not entirely a self-contained unit because the air which returns to the sea at upper levels spreads out and does not necessarily all return at the surface. The sea breeze rarely has a depth of more than 1,000 m and at its maximum strength does not extend inland more than 50 km. (The distance is much less around lakes, where the wind is more properly called a lake breeze.) It begins offshore in the late morning hours and gradually moves inland to decrease afternoon temperatures. Toward evening it subsides. Because of the sea breeze, areas on the immediate coast in middle latitudes and the tropics may have cooler temperatures than a few miles inland.

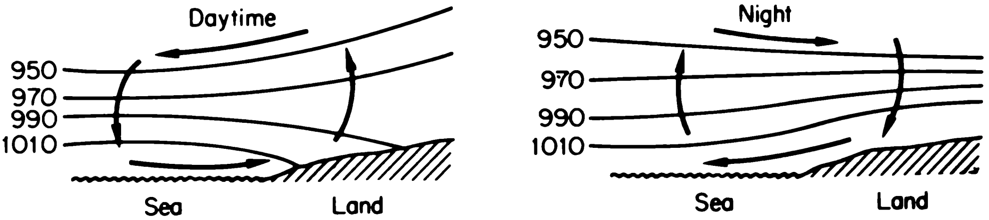


Figure 4.16
Principle of the sea and land breezes.

At night a reversal of the sea breeze may occur but with somewhat weaker characteristics. The *land breeze* becomes established during the night as the land cools to temperatures lower than the adjacent water, setting up a pressure gradient from land to sea (Figure 4.16). Because temperature differences between land and water are rarely as great at night as in the daytime, the land breeze is less extensive both vertically and horizontally than the sea breeze. It usually attains its maximum intensity in the early morning hours and dies out soon after sunup.

Another combination of local winds that undergoes a daily reversal consists of the *mountain* and *valley breezes.* On mountain sides on clear nights the higher land radiates heat and is cooled, in turn cooling the air in contact with it. The cool, denser air thus flows down the mountain slopes to drain into the valleys and lowlands. Since it blows from the mountain, this air flow is termed a *mountain breeze.* By morning it may produce temperature inversions in depressions so that the valley bottoms are colder than the hillsides. But, where the mountain breeze is funnelled into a narrow valley or a gorge, it may gain considerable velocity and generate enough turbulent mixing to break up the inversion. On warm sunny days, the heating of mountain slopes may generate an upslope flow of air called a *valley breeze.* As the warm air moves up the mountain, it is replaced by cooler air from above the valley, and surface temperatures are moderated slightly.

A type of wind known as a *gravity* or *katabatic wind* occurs in several parts of the world. Gravity winds result from the drainage of cool air off high plateaus or ice fields. The *mistral* flows onto the Mediterranean coast of France in winter from the higher lands and snow-capped mountains to the north. It is channelled somewhat by the Rhone Valley, and its coolness, dryness, and velocity sometimes detract from the otherwise attractive climate of the Riviera in winter. Along the northern coast of the Adriatic Sea, a cold, northeast wind known as the *bora* flows down from the plateau region in Yugoslavia onto the narrow coastal plain.

In numerous instances the winds associated with particular storm or pressure conditions occur with some regularity and are characteristic of the climate of a given locality. The Guinea Coast of Africa experiences the dry, dusty *harmattan* in winter when air from the relatively cool Sahara replaces humid tropical air. The harmattan has a marked cooling effect and is sometimes known as the *doctor.* In spring, as the subtropic high moves northward, the *sirocco* blows from the south across the Mediterranean lands. In Egypt, where it is known as the *khamsin,* and in other parts of North Africa, it is hot and dry. Along European coasts of the Mediterranean the sirocco has more moisture in its lower layers, but its high temperatures often have a withering effect on crops. *Leveche* is the name given to this wind in Spain. In the eastern Mediterranean northerly winds that persist at times during summer are known by the classical Greek name *etesians.*

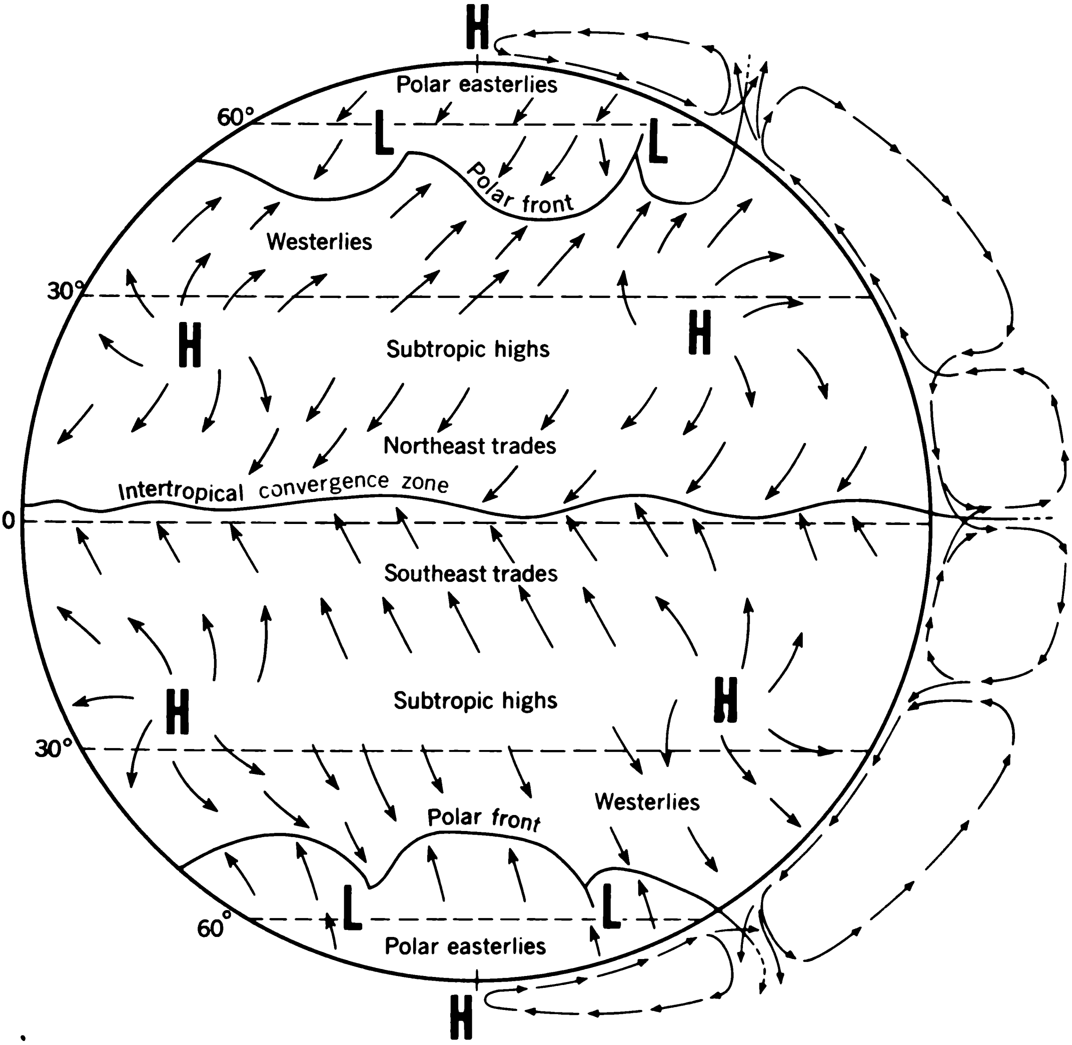
Winds that bring abrupt changes in temperature or moisture conditions are likely to take on local names. The *norther* marks the onset of cold, stormy weather in winter over Texas, the Gulf of Mexico, and the western Caribbean. The *pampero* is a squally wind from the northwest that blows over the Argentine pampas in winter. The *chinook* of North America, the *foehn* of the European Alps, and the *nor'wester* of New Zealand are warm, dry, gusty winds induced by mountain ranges. They will be discussed in more detail in connection with special storm effects in Chapter 5.

GENERAL ATMOSPHERIC CIRCULATION

The net effect of the differential heating of the earth by insolation is to produce density differences that set the atmosphere in three-dimensional motion. Much of the energy for maintaining the global circulation comes from the tropical oceans, where evaporation transfers large amounts of latent heat to the air. Although the influence of latitude upon heating might be expected to create a simple circulation between tropical and polar areas, the effect of the earth's rotation diverts the winds into gigantic whirling systems, or vortices, that are aligned more or less in latitudinal belts. The resulting zonal flow patterns have prevailing winds with strong easterly or westerly components and comprise the *general circulation.* If, for the time being, we neglect the influences resulting from differences in heat and moisture exchange over land and water surfaces we find a hypothetical arrangement of surface wind and pressure belts, as shown in Figure 4.17.

In the vicinity of the equator, where pressures are low, winds converge and rise, and the surface winds are light and variable. This *intertropical convergence zone,* also known as the *doldrums,* fluctuates in position and intensity and is at times a weak, discontinuous belt. On either side of the intertropical convergence, blowing into it (converging), are the *trades.* They are named the *northeast* and *southeast trades* in the Northern and Southern Hemispheres respectively. Note that although the pressure gradient is directed from the subtropic high toward the intertropical convergence the winds are deflected by the earth's rotation so that they approach the equator at acute angles rather than at the perpendicular. The sources of the trades are the subtropic highs, sometimes called the *horse latitudes,* where much of the upper air that is moving from the equatorial zone converges and piles up, then subsides and diverges near the surface. The subtropic highs are not actually continuous belts, but are broken up into "cells," having their best development over the oceans. Because much of the air movement is downward a tendency to calms or variable winds characterizes these cells. A part of the diverging air becomes the trades; that which flows toward the pole forms the *prevailing westerlies.* In the Northern Hemisphere the westerlies are from the southwest and in the Southern Hemisphere from the northwest. In each case the Coriolis effect accounts for the westerly component. The westerlies are zones of cyclonic storms, and although strong winds in these storms may blow from any direction of the compass, zonal westerlies predominate. The cyclonic storms themselves move in a general west to east direction. Land masses disrupt the westerlies considerably in the Northern Hemisphere, but in the Southern Hemisphere, where there is a virtually unbroken belt of water between 40° and 60° south, the westerlies are strong and persistent. They are often called the "roaring forties" in this zone, a carry-over from sailing days.

Figure 4.17
Schematic arrangement of winds and pressure in the general circulation. The effects of differential heating of land and water surfaces are neglected.



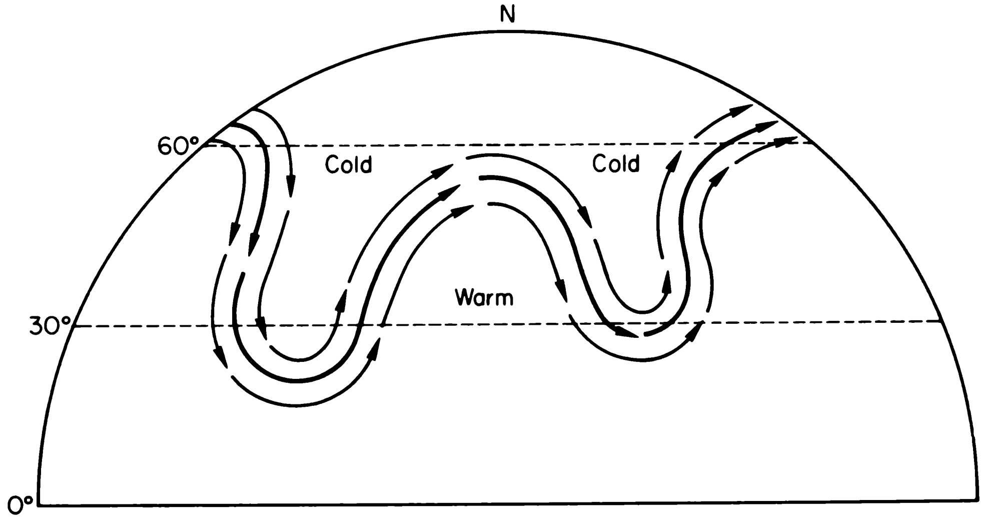
The westerlies and the *polar easterlies* meet and converge at the *subpolar lows,* or *polar fronts.* Here there is frequently a great contrast between the temperatures of the winds from subtropical and polar source regions, giving rise to the cyclonic vortices or "lows" that are carried along in the westerlies. The polar easterlies carry air outward from the *polar highs,* which are regions of subsidence of air from higher levels.

UPPER-LEVEL WAVES AND JET STREAMS

Completion of the circulation pattern of the major wind systems requires an exchange of air at the upper levels to compensate for any net transfer horizontally at the bottom of the troposphere. Strong air streams in the upper troposphere, lower stratosphere, and possibly even in the mesosphere help to achieve a balance in the global circulation. At the higher levels the mean flow is more nearly along parallels of latitude than is the case at the surface. Nevertheless, there is a slight poleward flow in latitudes 0° to 30° and 60° to 90° and an equatorward flow in the middle latitudes. The mean vertical motion that completes the three-dimensional circulation is upward in the intertropical convergence zone and along the subpolar low; it is downward in the subtropical high-pressure cells and near the poles. These vertical transfers are of small magnitude compared to horizontal motion, but they are quite essential for maintenance of the general circulation and development of storms.

The winds at several kilometers above the surface follow giant, undulating paths around the earth in the latitudes of the westerlies (Figure 4.18). These waves apparently result from the tendency of winds

Figure 4.18
Wave trajectories in the upper westerlies of the Northern Hemisphere. Heavy arrows represent typical jet-stream paths.



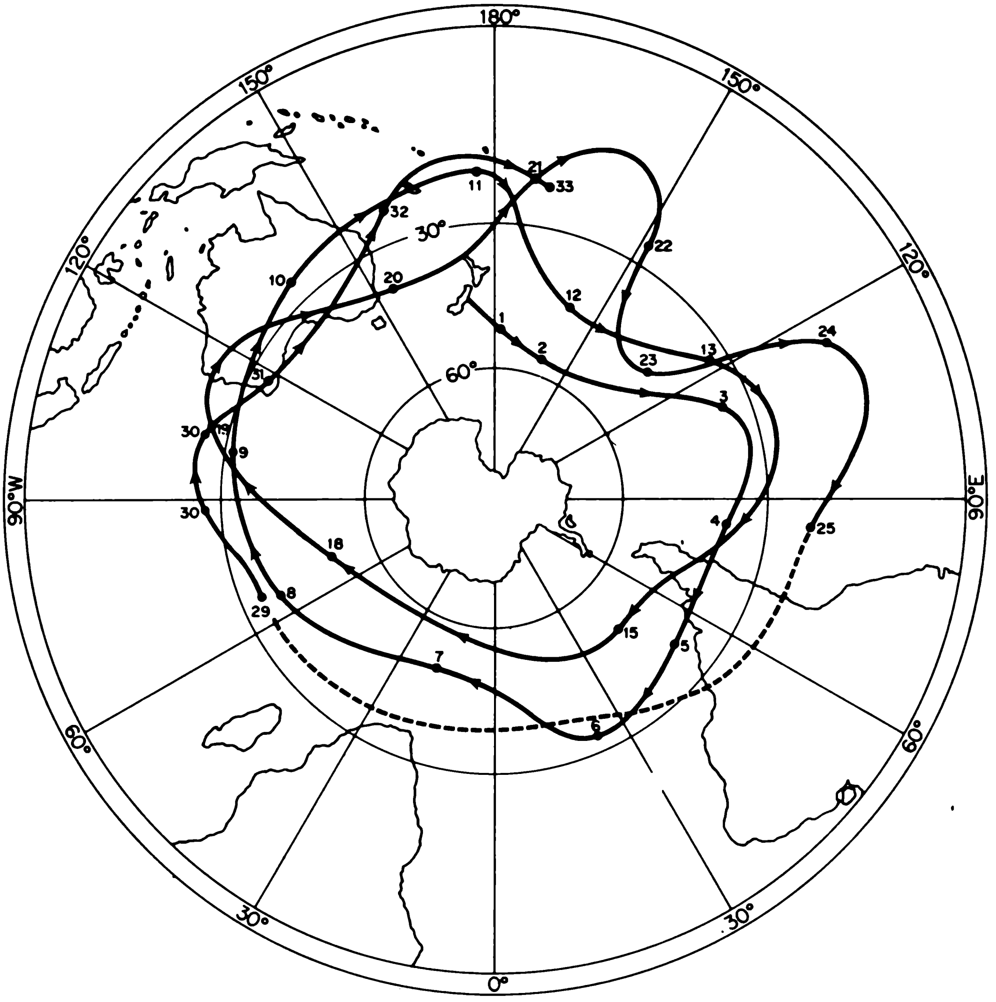


Figure 4.19
Path and daily positions of a constant-level balloon in the Southern Hemisphere upper westerlies. After launching at Christchurch, New Zealand, on March 30, 1966, the GHOST balloon drifted at an altitude of about 12 km in upper-level waves for 33 days. (NCAR [National Center for Atmospheric Research], Boulder, Colorado)

in large-scale motion systems to retain a constant spin, or angular momentum, about the earth's axis of rotation. A stream of air moving toward the equator adopts a cyclonic curvature (counterclockwise in the Northern Hemisphere) relative to the surface at the lower latitude as the distance from the axis of rotation increases. Eventually the curved path turns the wind back toward the pole. Passing its original latitude, the wind takes an opposite (that is, anticyclonic) curvature relative to the earth as it comes closer to the polar axis, where the rotational velocity is less. (These effects are easily demonstrated on a globe.) The resulting *Rossby waves* may have lengths of 3,000 to 6,000 km, permitting three to six circumpolar waves that correspond to surface paths of depressions and anticyclones in the westerlies (see Figure 4.19). Their length, amplitude, and position are influenced by differential heating at the surface and by ex-

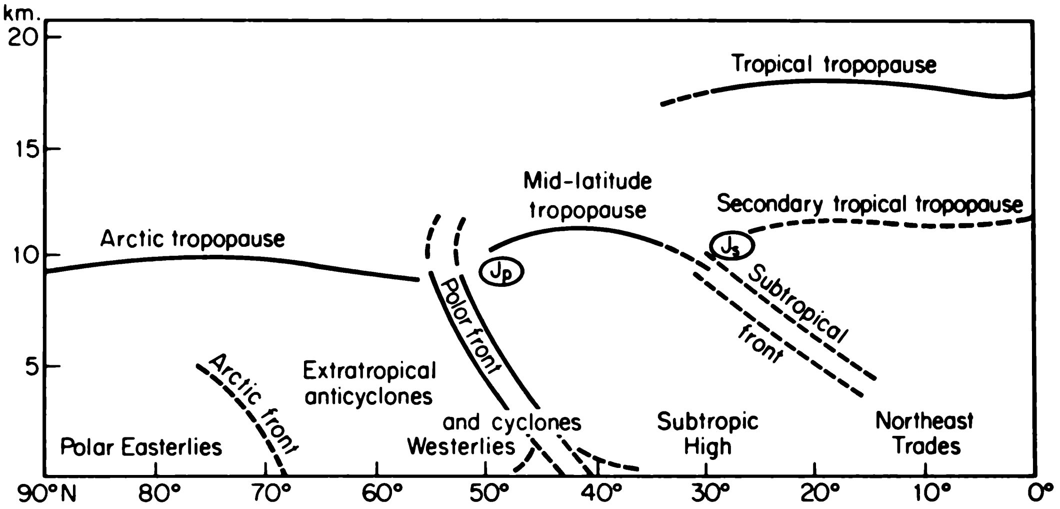


Figure 4.20

Surface wind and pressure systems in relation to the fronts and jet streams of the Northern Hemisphere. Dashed lines represent zones of major air mass exchange. (After Defant and Taba, Newton and Persson, and Palmén)

tensive mountain barriers. Major waves tend to persist in the westerlies above the Rockies, the Andes, and the plateaus of Central Asia and South Africa. Long waves often remain stationary for considerable periods, although winds within the wave pattern move at great speeds.

Winds in upper-level tropospheric waves reach maximum speeds in the *jet streams,* narrow bands of high-velocity winds that follow the wave path near the tropopause at elevations of 8 to 15 km. The ***polar front*** *jet stream* achieves its maximum force and extent in winter, when there may be two or even three distinct currents having wind speeds of 100 knots or more at their cores. In the Northern Hemisphere the main jet stream system undulates from north to south as part of the upper-level westerly wave, often swerving far equatorward over the continents. Sometimes it divides into apparently discontinuous segments. Its position relative to the polar front suggests that it may be a guiding mechanism for cyclonic storms as well as a means of transporting great quantities of air across the continents (see Figure 4.20). In summer the polar front jet has a mean position at higher latitudes, and its velocity and extent are reduced. A similar jet stream system in comparable latitudes of the Southern Hemisphere crosses South America, Australia, and New Zealand.

Above the subtropic highs a high-speed westerly flow, the ***subtropical jet stream,*** persists through most of the year. Its wind speeds commonly exceed 100 knots; maximums of more than 300 knots have been recorded in both hemispheres, for example, over Japan and the South Indian Ocean.

In the vicinity of the equator upper-level wave motion is not so well developed as at higher latitudes owing to a negligible Coriolis effect and a relatively steady flow of air that does not push vigorously across parallels of latitude. Nevertheless, in the tropical stratosphere an *easterly jet stream* develops south of Asia during the Northern Hemisphere summer. Although its direction is steady, it varies in speed, averaging about 70 knots. The tropical easterly jet is thought to be associated with thermal conditions arising from summer heating of the Asian continent. Also known as the Krakatoa easterlies, it probably carried volcanic dust westward after the eruption of Krakatoa in 1883.

Another stratospheric jet stream is the *polar night jet.* During winter the sun does not heat a cone of atmosphere at the pole, and the resulting strong temperature gradient in the stratosphere creates westerly winds at heights averaging 60 km. Rocket observations have recorded speeds exceeding 330 knots at a height of nearly 80 km in the polar night jet stream above the Arctic.

SEASONAL CHANGES IN THE GENERAL CIRCULATION

The global patterns of temperature, pressure, and circulation migrate seasonally along the meridians. Figure 4.21 shows contrasting seasonal positions of the intertropical convergence. The shift is basically the result of seasonal changes in the heat budget at different latitudes. Not only is there a continual change in the declination of the sun, but continents and oceans also assume different roles in response to insolation. The tendency is for wind and pressure belts to shift northward a few degrees in the Northern Hemisphere summer and southward in winter. It should be emphasized, therefore, that the distribution of wind and pressure systems in the general circulation, outlined earlier in this chapter, represents idealized pressure distribution and prevailing air movements that

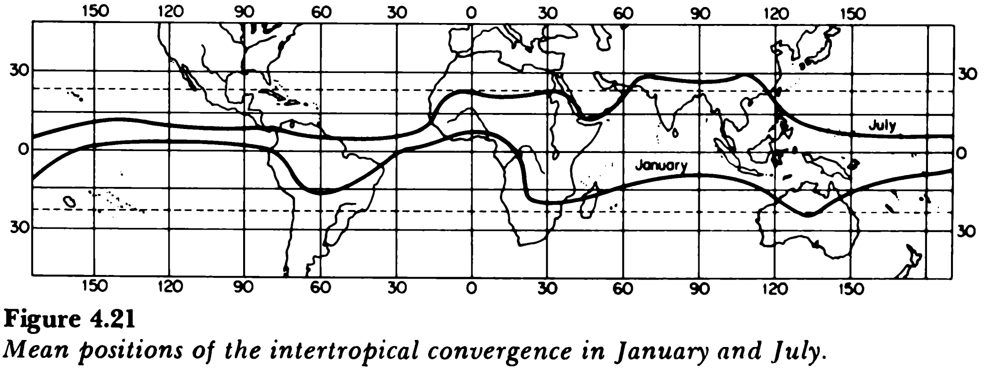


Figure 4.21
Mean positions of the intertropical convergence in January and July.

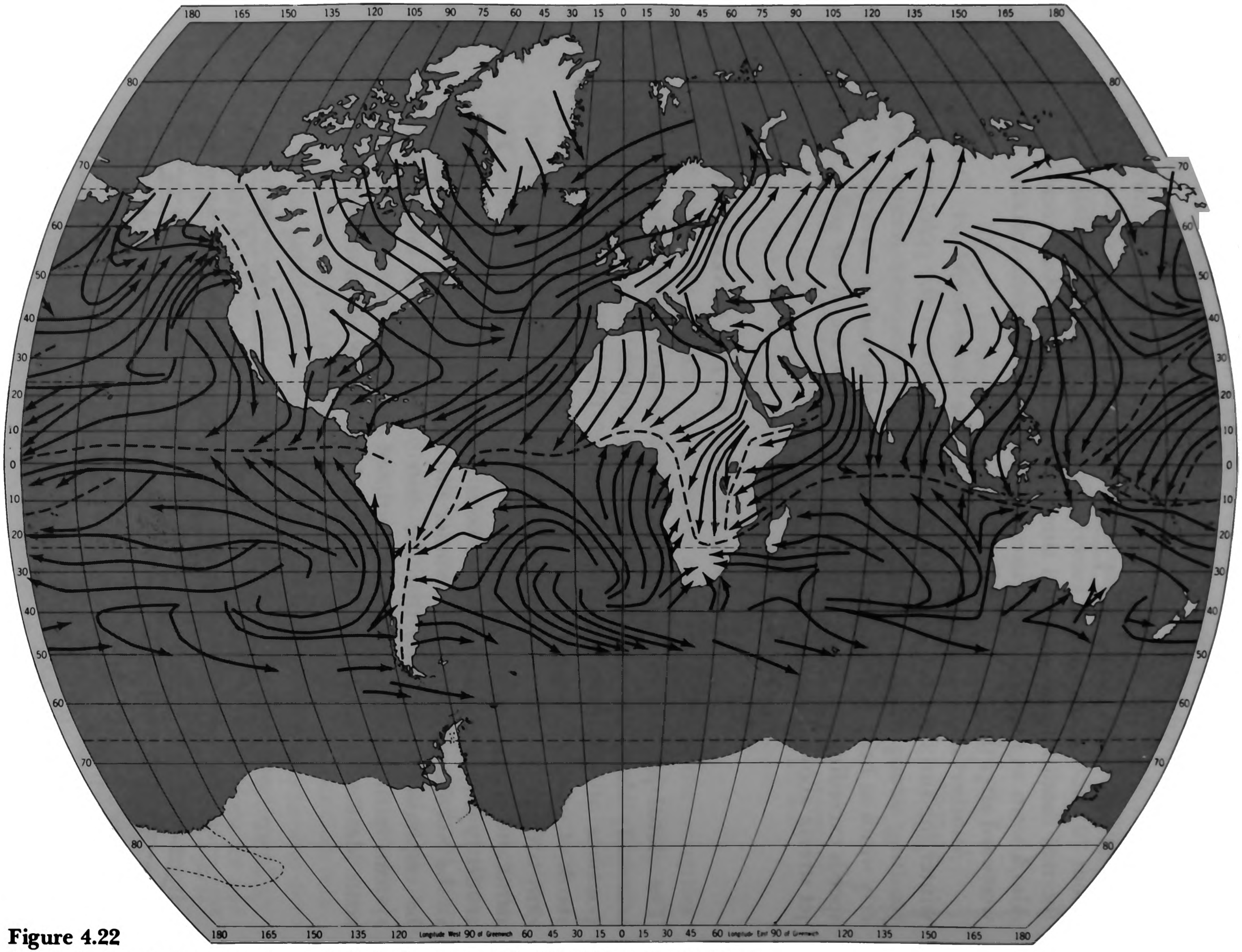


Figure 4.22
Mean flow patterns of surface winds in January. (Modified Van der Grinten projection)

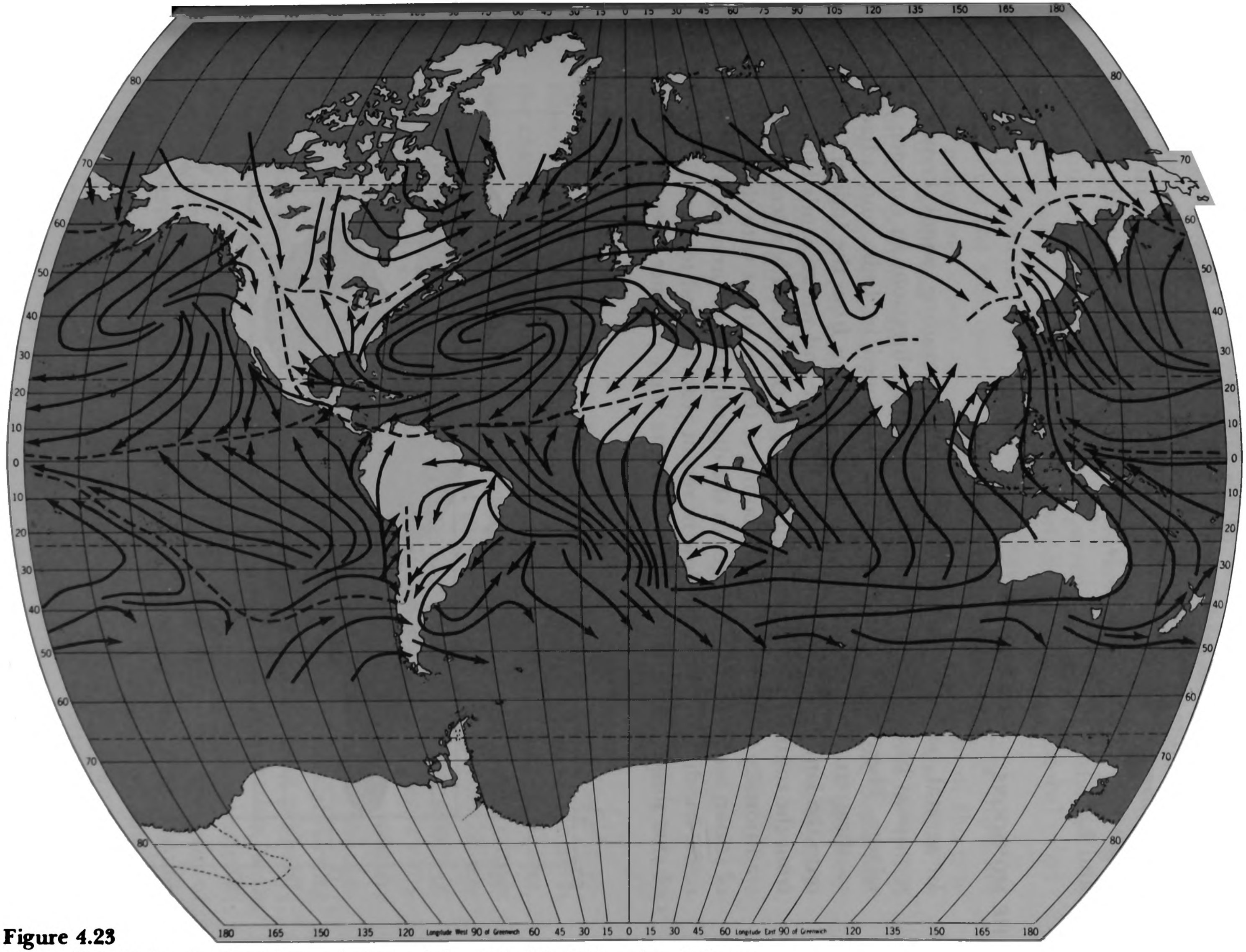

Longitude West 90 of Greenwich
Longitude East 90 of Greenwich
180
165
150
135
120
60
45
30
15
0
80
70
60
50
40
30
20
10

Figure 4.23
Mean flow patterns of surface winds in July. (Modified Van der Grinten projection)

are modified by several factors. It has been previously pointed out that pressures tend to be higher in winter over land and lower over adjacent waters. This is most evident in the middle latitudes. The result may be creation of complete reversals of pressure gradients over the continents with corresponding seasonal shifts in wind direction. The mean flow patterns of surface winds in January and July are shown in Figures 4.22 and 4.23.

THE MONSOONS

In several parts of the world seasonally prevailing winds known as *monsoons* blow from approximately opposite directions in summer and winter. These wind systems are best developed in India and adjacent southeastern Asia, where there is a persistent flow from the Indian Ocean onto the land in the summer half of the year and where winds blow out from the continent in winter. This reversing circulation has been ascribed traditionally to the heating and cooling of the continent with the changes of season and resulting changes in the distribution of pressure. Increased understanding of upper-air flow in recent years has led to a modification of this simple explanation. In winter the high pressure over the Asian

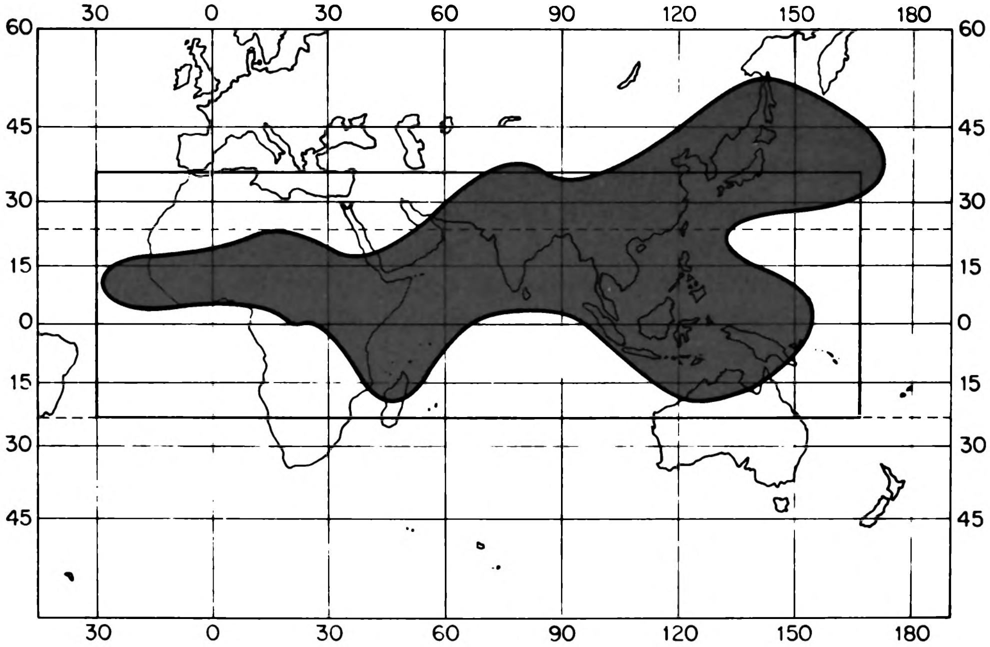


Figure 4.24
The monsoon region. (Shaded area after Khromov; rectangle after Ramage)

continent forces the westerly jet stream southward so that a portion of it lies south of the Himalayas. Subsiding air along the tropical margins of the jet moves toward the Indian Ocean, blocking incursions of maritime tropical air and producing the dry winter monsoon. In summer the jet stream shifts with the general circulation and lies north of the Himalayas, allowing moist tropical air of the wet summer monsoon to move onto the continent. Northern Australia experiences a similar, though less well-developed, monsoonal circulation which is related to seasonal migration of components of the general circulation.

Along the east coast of Asia and across Central Africa there are other monsoonal effects that merge into the shifting pattern of the general circulation (see Figure 4.24). Outbursts of polar air in the westerlies result in a predominantly westerly flow off northeastern Asia in winter, but in summer maritime air moves inland from the subtropic highs that lie over the adjacent oceans. In Africa the alternating flow arises from seasonal changes in heating and pressure over the deserts, and it is modified by temperatures over the adjacent seas.

OCEANIC CIRCULATION

There is a close relationship between the primary circulation of the atmosphere and the circulation of the oceans. Although the atmosphere is the chief agent for transfer of energy from the tropics to higher latitudes, the oceans also serve this function on a global scale. In addition, energy is exchanged between the oceans and the atmosphere by transfer of sensible and latent heat and by mechanical action along the ocean-air boundary surface. Ocean currents are motivated mainly by density differences due to variations in water temperature and salinity and by winds. Their direction is influenced by the Coriolis effect and by the configuration of ocean basins and shorelines as well as by the initial driving forces. Large-scale, slow movements of ocean water, resulting primarily from prevailing winds, are usually referred to as *drifts* rather than currents. These are best developed in the belt of prevailing westerlies in the Southern Hemisphere (see Figure 4.25). Examination of the map of world ocean currents reveals the close correlation between general atmospheric circulation and oceanic circulation in the middle and low latitudes. In the Northern Hemisphere the circulation is clockwise; in the Southern Hemisphere it is counterclockwise. The net effect of most of the surface oceanic circulation is to carry cold water toward the equator along the east margins of the oceans and warm water toward the poles along the west margins. At high latitudes the pattern of flow is broken up

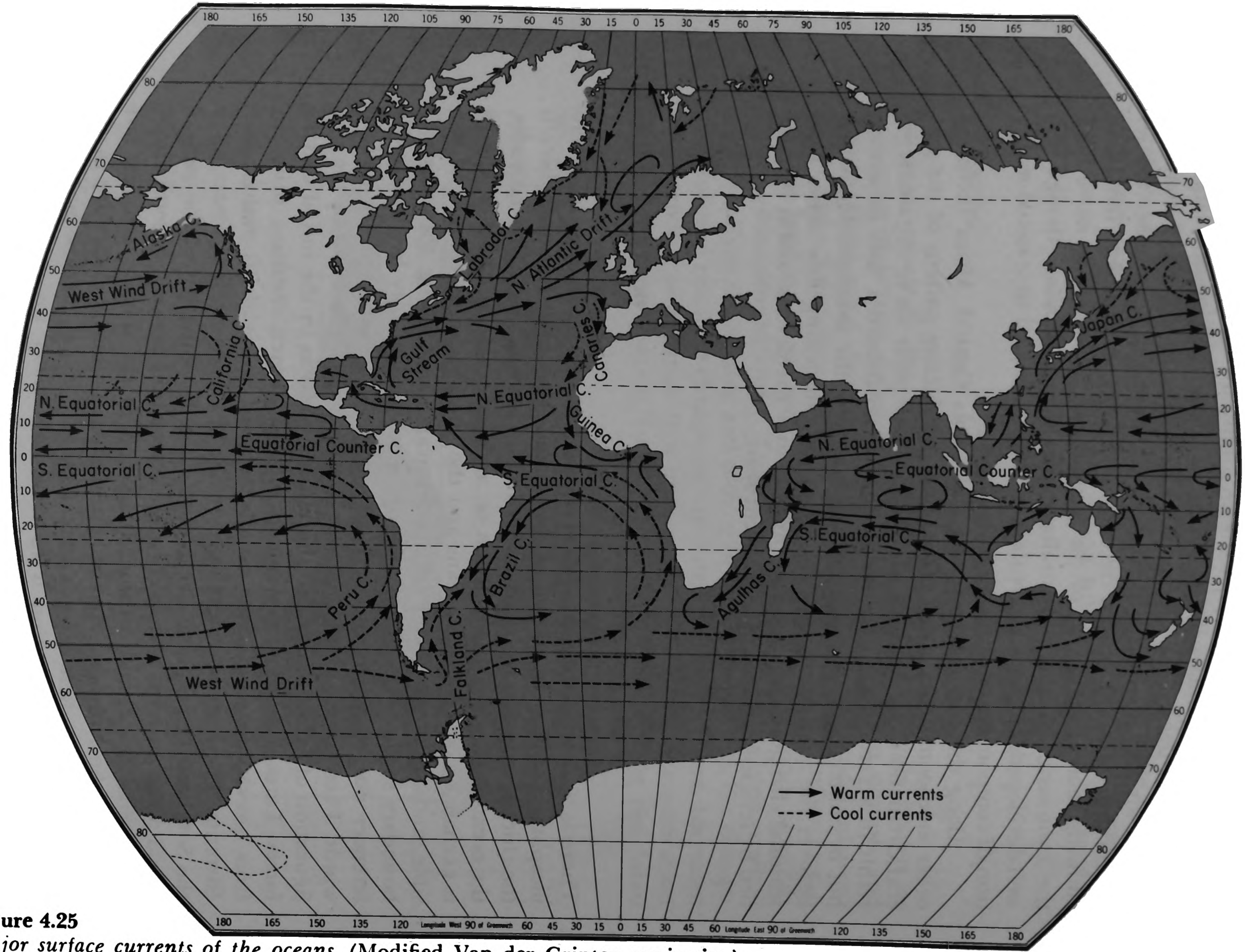


Figure 4.25
Major surface currents of the oceans. (Modified Van der Grinten projection)

considerably by the arrangement of land masses. Cold water subsides in some of these areas and moves toward the equator at great depths.

Along some coasts prevailing winds skim the surface water away, producing a phenomenon known as *upwelling*. When the relatively warmer surface water is moved out by the action of wind the cooler water from below replaces it. Upwelling is especially important in influencing temperatures along the coasts of southern California, Peru and northern Chile, northwest Africa, and southwest Africa. In each of these areas the Coriolis effect turns the currents away from the shore. Since the surface water has a greater speed than that at greater depths, it is deflected at a greater rate, allowing upwelling of the colder subsurface water. As a result of the abnormally low temperatures at surface levels, moist air passing over the areas of upwelling frequently produces fog.

Additional Readings

Batten, E. S., "Wind Systems in the Mesosphere and Lower Ionosphere," *Journ. of Met.,* 18, 3 (1961), 283–91.

Hidy, George M., *The Winds.* Princeton, N.J.: D. Van Nostrand Co., Inc., 1967.

Lamb, H. H., "The Southern Westerlies," *Quart. Journ. Roy. Met. Soc.,* 85, 363 (1959), 1–23.

Lorenz, Edward N., "The Circulation of the Atmosphere," *Am. Scientist,* 54, 4 (1966), 402–20.

Ludlum, David M., "Extremes of Atmospheric Pressure in the United States," *Weatherwise,* 15, 3 (1962), 106–15.

McDonald, James E., "The Coriolis Effect," *Scient. Am.,* 186, 5 (1952), 72–78.

Namais, Jerome, "Interactions of Circulation and Weather Between Hemispheres," *Mon. Wea. Rev.,* 91, 10–12 (1963), 482–86.

Newell, Reginald E., "The Circulation of the Upper Atmosphere," *Scient. Am.,* 210, 3 (1964), 62–74.

Palmén, E., and C. W. Newton, *Atmospheric Circulation Systems.* New York: Academic Press Inc., 1969.

Panofsky, H. A., "Up and Down," *Weatherwise,* 25, 2 (1972), 77–82.

Reiter, Elmar R., *Jet Streams.* Garden City, N.Y.: Doubleday & Company, Inc., 1967.

Serguis, Leo A., George R. Ellis, and Richard M. Ogden, "The Santa Ana Winds of Southern California," *Weatherwise,* 15, 3 (1962), 102–5.

chapter five

Weather Disturbances

Travelling along as part of the large-scale motion systems of the general circulation are the air masses and storms that bring our daily weather and, by their cumulative effect, produce different climates. Most major weather changes consist of the advances and interaction of air masses and of processes within the air masses themselves. Mass transport of moisture from the oceans accounts for much of the water that later becomes precipitation over the continents, and air masses in motion are largely responsible for the transfer of temperature properties and latent heat from one region to another. The analysis of atmospheric conditions over an area must include air masses and storms as well as the individual elements of weather that comprise those conditions.

PROPERTIES OF AIR MASSES

An *air mass* is defined as an extensive portion of the atmosphere having characteristics of temperature and moisture which are relatively homogeneous horizontally. For a large body of air to acquire temperature and moisture properties that are approximately the same at a given level, that air must rest for a time on a *source region,* which must itself have fairly homogeneous surface conditions. A large land or water area which has evenly distributed insolation affords a good source region, but a second prerequisite is necessary if a distinctive air mass is to be developed, namely, large-scale subsidence and divergence of air over the source region. Air that subsides over a homogeneous source region will become homogeneous itself and tends to retain its characteristics when it moves away. The heat and moisture properties of the air mass gradually change, however, as it moves over other surface conditions. Zones of convergence and rising air are inimical to the production of air masses because the general movement of winds is toward these areas at surface levels, bringing a constant renewal of air with heterogeneous temperature and humidity properties.

AIR MASS IDENTIFICATION AND ANALYSIS

Once an air mass has left its source region, it undergoes changes which often make it difficult to identify by ground observers. For example, if it is formed over a cold surface and subsequently passes over a warm ocean its temperature and moisture content will increase. Local surface conditions created by ocean currents, land relief, minor water bodies, or nighttime radiation can produce quite different values of temperature and humidity at the bottom of an air mass. Consequently, it is necessary to analyze conditions in the upper air and to understand the processes that accompany the change in air mass properties.

Upper-air observations to provide information for air mass analysis are carried out by means of the *radiosonde* (Figure 5.1). For the sake of convenience, radiosonde observations are called *raobs.* The radiosonde consists of a lightweight box fitted with a radio transmitter and sensing devices for pressure, temperature, and relative humidity. Although there are variations in the construction of radiosondes, they commonly employ a sylphon cell to indicate pressure, a thermistor for temperature, and a coated plastic strip for relative humidity. The instrument package rides aloft under a gas-filled balloon and descends on a parachute after the balloon bursts at an altitude of about 25 to 30 km. At a ground station the signals are received, recorded, and translated into usable data. Analysis of raobs from a series of stations enables the forecaster to determine the kinds of air masses that prevail over a region. At some stations the radiosonde is followed by radar to obtain wind direction and speed at various levels as in the case of rawin. The combined technique is known as *rawinsonde.*

Another type of airborne instrument, the *transosonde,* is a balloon-borne radiometeorological device whose altitude can be controlled by

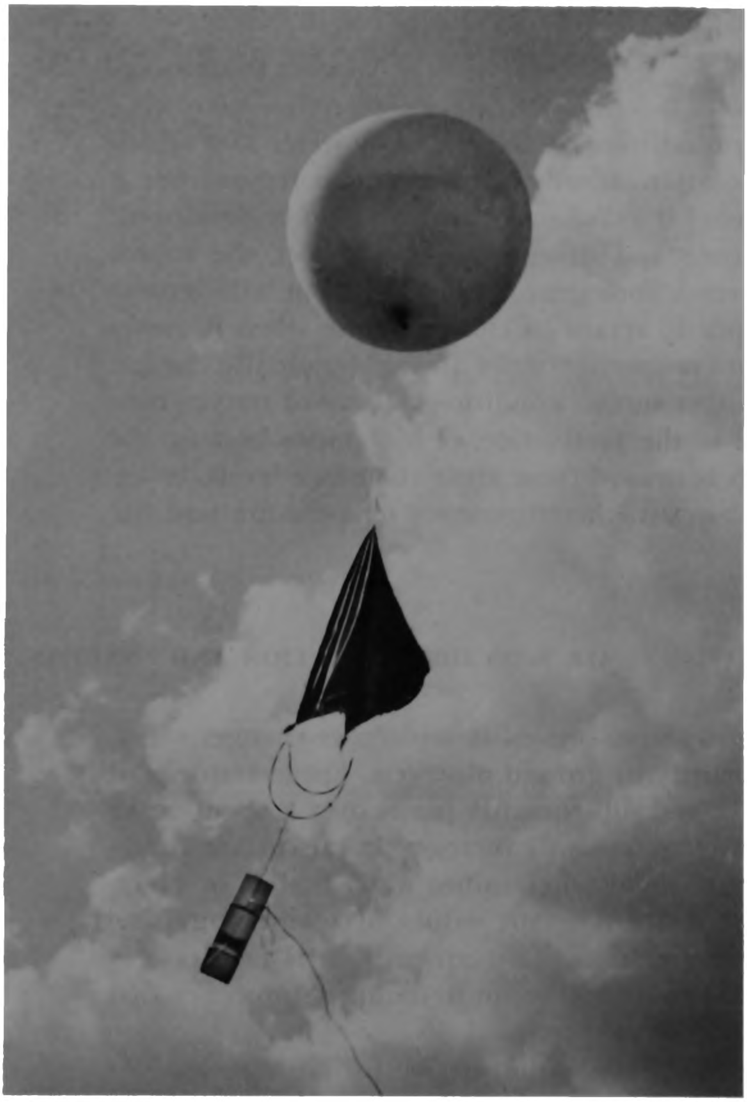


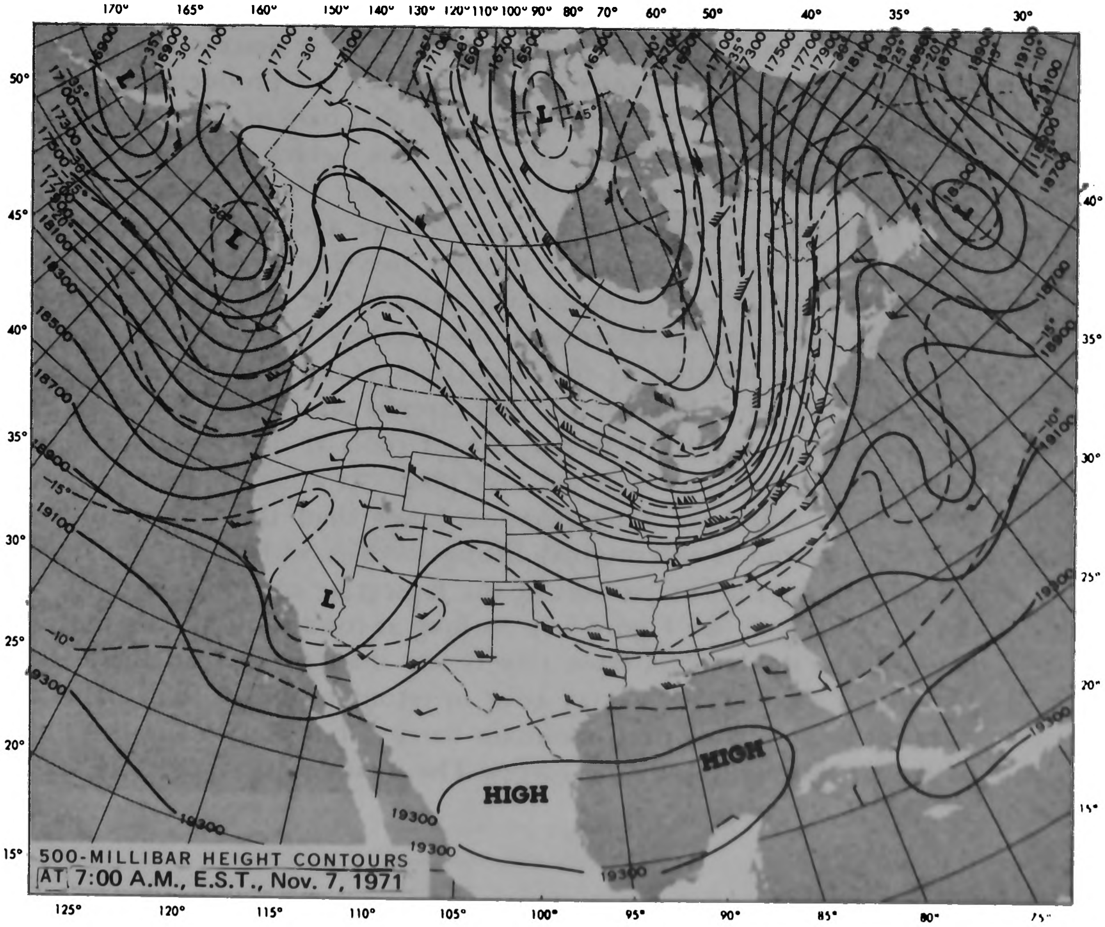
Figure 5.1
Radiosonde in flight.
(The Bendix Corporation, Environmental Science Division)

releasing ballast or gas from the balloon as it floats along at a constant pressure level for several days. Its advantage is that it can obtain data from positions over the oceans, where other sources of information are limited. Giant adaptations known as *GHOST* (Global Horizontal Sounding Technique) balloons have provided valuable data from upper levels, especially in the Southern Hemisphere westerlies. Unfortunately, transosonde balloons create a hazard for high-altitude jet aircraft.

Identification of air masses is based upon three kinds of information: (1) the history of change in the air since it left its source region; (2) horizontal characteristics at certain levels in the upper air; and (3) the vertical distribution of temperature, winds, and humidity. Because the number of raob reporting stations is limited and because major changes in air masses may occur between the times of radiosonde observations, the historical approach alone is inadequate. Maps plotted with temperature, wind, and humidity values for one or more upper levels have the advantage of showing a minimum of the influences due to local surface conditions and of revealing moisture and energy components of the upper part of the air mass. They represent a horizontal "slice" through the atmosphere. Constant-level charts may be plotted either for a certain height or for a constant-pressure level, the latter being most often used by meteorological services (Figure 5.2). Common pressure levels for constant-pressure analyses are 850, 700, 500, 300, and 100 mb.

Just as charts showing the horizontal distribution of weather elements in the upper atmosphere help to identify air masses and their boundaries, so vertical cross sections based upon raob and upper wind reports make it possible to analyze vertical differences in the air. Wind direction and velocity in particular are useful in judging air mass movements and in flight planning. Temperature and humidity values from the raob are analyzed as a basis for forecasting temperature changes, cloudiness, storms, and precipitation.

Figure 5.2
A 500-mb constant-pressure chart for North America. Radiosondes gather data for maps of this type. Solid lines are contours indicating the height of the 500-mb surface above sea level. Dashed lines are isotherms for intervals of 5°C. Wind symbols show relative wind speed at the 500-mb level. (NOAA, Environmental Data Service)



STABILITY AND INSTABILITY

Vertical movements of air are of great significance because nearly all precipitation is associated with adiabatic cooling and condensation in rising air. The fundamental causes of rising air have been discussed. Let us now examine the factors within the air which affect its tendency to vertical motion, that is, its stability. If a mass of air tends to remain in its position, or to return to that position when displaced, it is said to be *stable.* If vertical displacement results in a tendency to further movement of the air from its original position, the air mass is termed *unstable.* If conditions in the air are such that it neither resists vertical motion nor aids it, the air mass is in a state of *neutral equilibrium.*

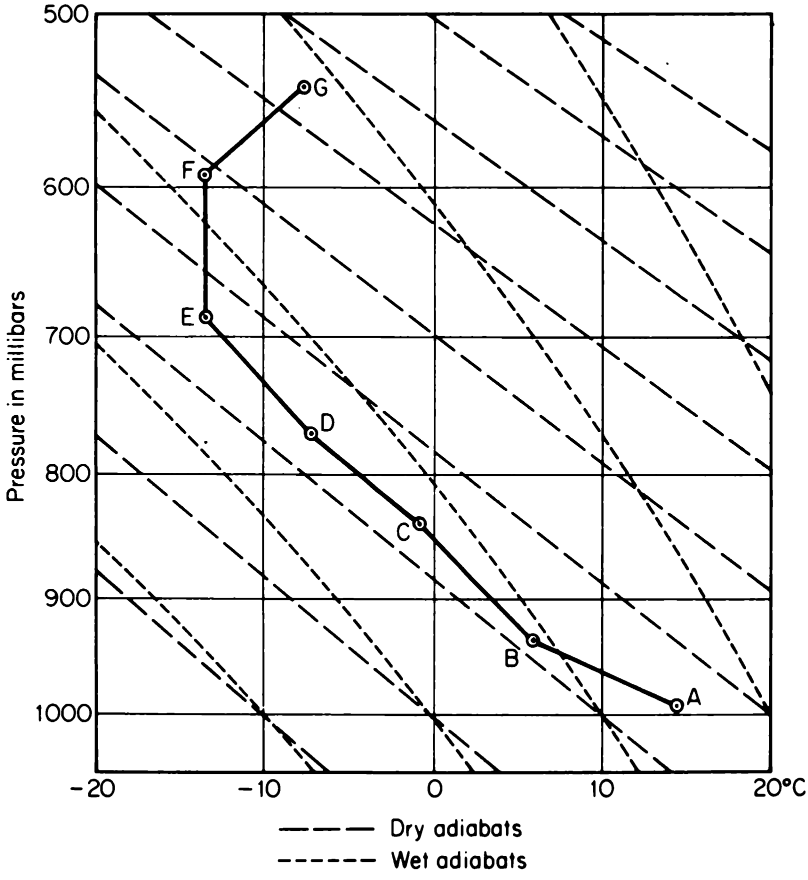
An air mass having colder, drier air in its surface layers than aloft is likely to be stable, and any process which cools the air at the bottom of an air mass tends to make it more stable. Radiational cooling at night, for example, produces a stable condition, that is, a temperature inversion. Advection of air over a cold surface will result in increasing stability, as when warm oceanic air passes over a cold ocean current or onto cold land. While these processes can produce condensation (fog or dew) in the lower layers of an air mass, they do not induce vertical currents and therefore are not likely to produce precipitation. Subsidence of air also increases its stability. As the bottom of a layer of subsiding air is compressed and diverges the top of the layer descends farther and is warmed more adiabatically than the base, thus producing a smaller difference in temperature between the top and base of the layer, that is, a decreased lapse rate. Upper-air inversions, which exhibit extremely stable lapse rates, are often caused by subsidence.

The vertical lapse rates of temperature and humidity determine the degree of stability of an air column (see Figure 5.3). If the actual temperature of the air increases with altitude, the colder, denser air below will tend to stay in place with the warmer, lighter air above—obviously a stable condition. If the actual temperature is exactly the same vertically through a layer of air, that layer is also stable. When an outside force causes stable air to rise, the air cools at the try adiabatic rate as long as condensation does not take place. At a higher altitude (lower pressure), it will have a temperature lower than the surrounding air and will descend if the original lifting force is withdrawn. This effect is best seen by the comparison of the lapse rate of stable air with the dry adiabatic rate (see Figure 5.3). As long as condensation does not take place, air is stable if its lapse rate is less than the dry adiabatic rate. Because of its tendency to stay at or to return to its position, stable air does not move readily over mountain barriers but flows along them and pours out through passes and gorges. The weather associated with stable air is generally fair, but visibility is often poor owing to ground fog or concentration of smoke and haze near the surface. In winter stable air masses may be extremely cold in middle and high latitudes.

Instability in air occurs when its lapse rate is greater than the dry adiabatic rate of cooling. If air in an unstable layer is forced aloft it will be warmer than the surrounding air and will continue to rise. The initial force may result from heating at the earth's surface, from orographic lifting, or from convergence. As long as the actual lapse rate of temperature is greater than the dry adiabatic rate, the air will continue to rise and cool adiabatically. When its saturation temperature is reached, condensation produces clouds, and precipitation may result. Warm, humid air masses are unstable, especially when they lie over a warm surface. Instability may be developed in any air mass if it is warmed and if its moisture content is increased in its lower layers.

It frequently happens that air is stable while it rises and cools at the dry rate (that is, without condensation) but becomes unstable when

Figure 5.3
Hypothetical lapse rate of temperature showing different stability conditions. AB, *superadiabatic, absolutely unstable;* BC, *conditionally unstable;* CD, *neutral equilibrium (unsaturated air);* DE, *neutral equilibrium (saturated air);* EF, *isothermal lapse rate, absolutely stable;* FG, *inversion, absolutely stable.*



condensation begins. Consider a mass of mildly stable air being forced aloft. It cools at the dry adiabatic rate until it reaches saturation (see Figure 5.3). The release of the latent heat of condensation gives added heat to produce further lifting and the air then cools at the slower wet rate. Strong convective activity and heavy precipitation is the probable result. Humid air which is initially stable but which becomes unstable when condensation takes place within it is termed conditionally unstable. *Conditional instability* can be identified by the fact that the actual lapse rate in the air is less than the dry rate and greater than the wet rate. That is, it is stable with respect to the dry rate but unstable with respect to the wet rate. Instability is thus "conditional" upon the presence of considerable water vapor in the air.

It should be apparent at this point that *neutral equilibrium* occurs when the actual lapse rate in dry air equals the dry adiabatic rate. Air rising and cooling at the dry rate has the same temperature as the surrounding air at every level. In saturated air neutral equilibrium is achieved when the lapse rate equals the wet adiabatic rate. Again vertical displacement results in the same temperature as that of the surrounding air. In both cases the conditions within the air produce a tendency neither to sink nor to rise.

In view of their importance in connection with vertical motion in air masses, stability and instability are significant properties for air mass analysis and forecasting. By studying the temperature and humidity values for different levels the meteorologist can ascertain the degree of stability and the associated potential weather of an air mass. The lapse rate as plotted from raob data is a profile of the upper air; the vertical temperature and humidity gradients in various layers of air are fundamental clues to stability conditions.

AIR MASS SOURCE REGIONS AND CLASSIFICATION

An air mass source region has already been defined as a large area with approximately homogeneous temperature and moisture properties where there is a general subsidence and divergence of air. These conditions are found to be best developed in the semipermanent high-pressure belts of the earth. In the belt of low pressure along the equator, however, the equatorial convergence may be weakly defined and stagnation of air will produce equatorial air masses. Where development of high pressure is seasonal, for example, over mid-latitude continental regions in winter, the source regions will likewise have seasonal maximum development.

Classification of air masses is based primarily upon their source regions and secondarily upon temperature and moisture properties. The two main categories are tropical or subtropical and polar or subpolar,

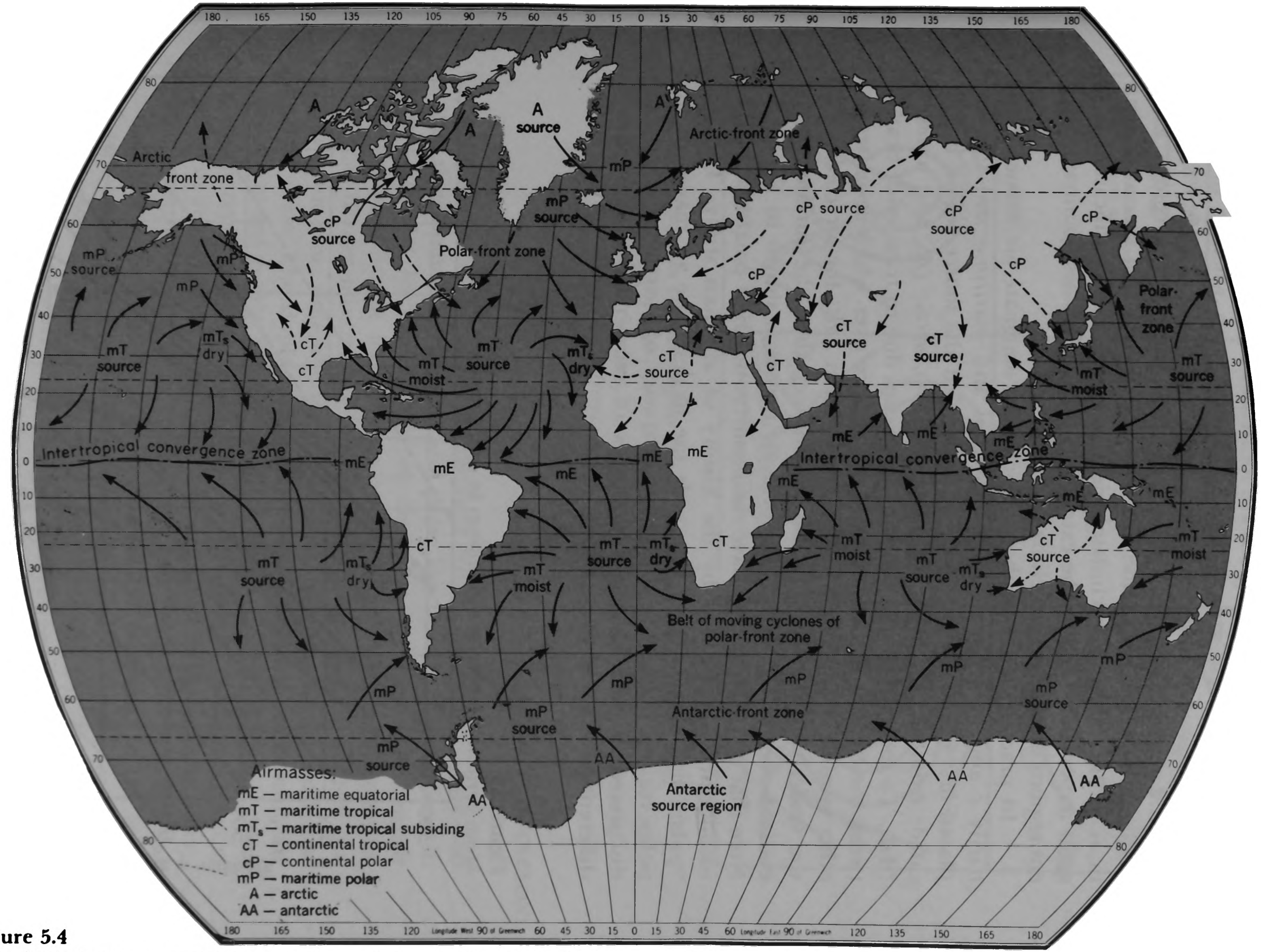
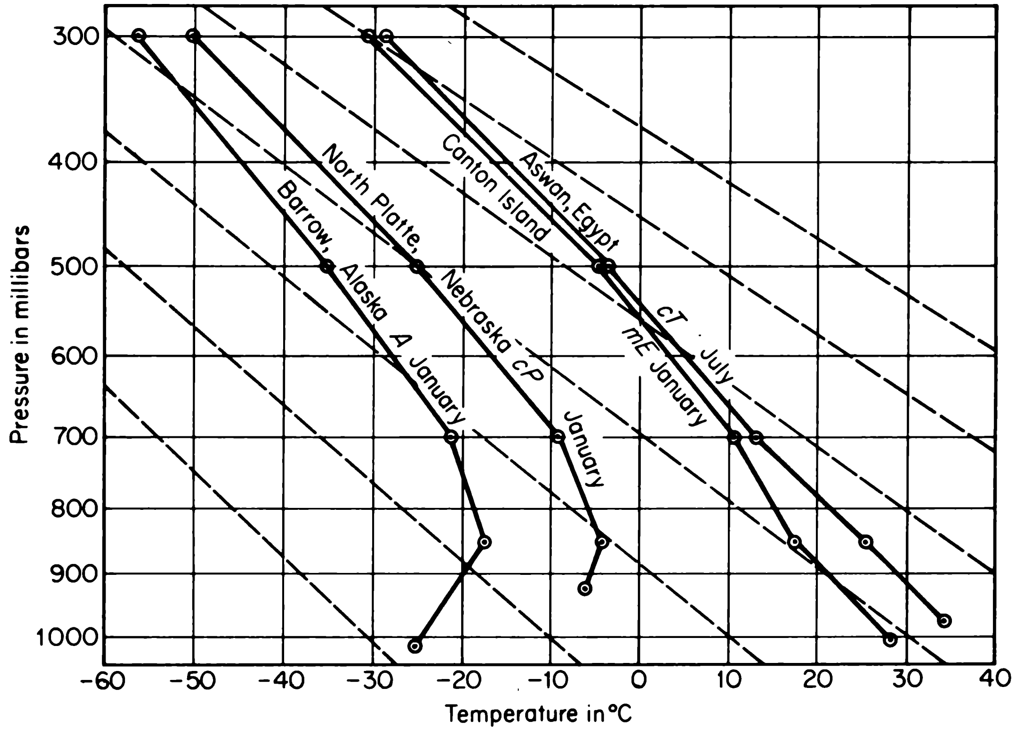


Figure 5.4
Principal air masses and source regions of the world. (Data after Strahler, courtesy John Wiley & Sons, Inc.)

because the great source regions are located at high and at low latitudes. Subdivision of these groups is made according to whether the source region is oceanic or continental, and further, according to what modifications the masses experience as they move from their source regions. Eventually, air masses become modified to such an extent that special designations are necessary. For climatological purposes this classification is of great importance because the extent to which air masses dominate different regions determines the climate of those regions.

In practice, letter symbols are used to designate air masses. Ordinarily *c* and *m* are used for *continental* and *maritime* and they are placed first in the designation. Following that the source region is indicated: *tropical* (*T*), *polar* (*P*), *equatorial* (*E*), *arctic* (*A*), and *antarctic* (*AA*). To indicate modifications of air masses due to transfer of heat between the bottom of the mass and the surface over which it passes, another symbol is appended: *k* (for the German *kalt*) for air colder than the underlying surface or *w* for air warmer than the surface. A generalized map of the principal world air masses is shown in Figure 5.4. Note that arctic air masses have their sources north of polar masses. This incongruity is explained by the history of air mass study. The term "polar" had already come into wide use to designate air masses in subpolar regions before the distinctive character of "arctic" air had been discovered. Figure 5.5 shows examples of temperature lapse rates in different air masses.

Figure 5.5
Mean vertical temperature soundings in contrasting air masses.



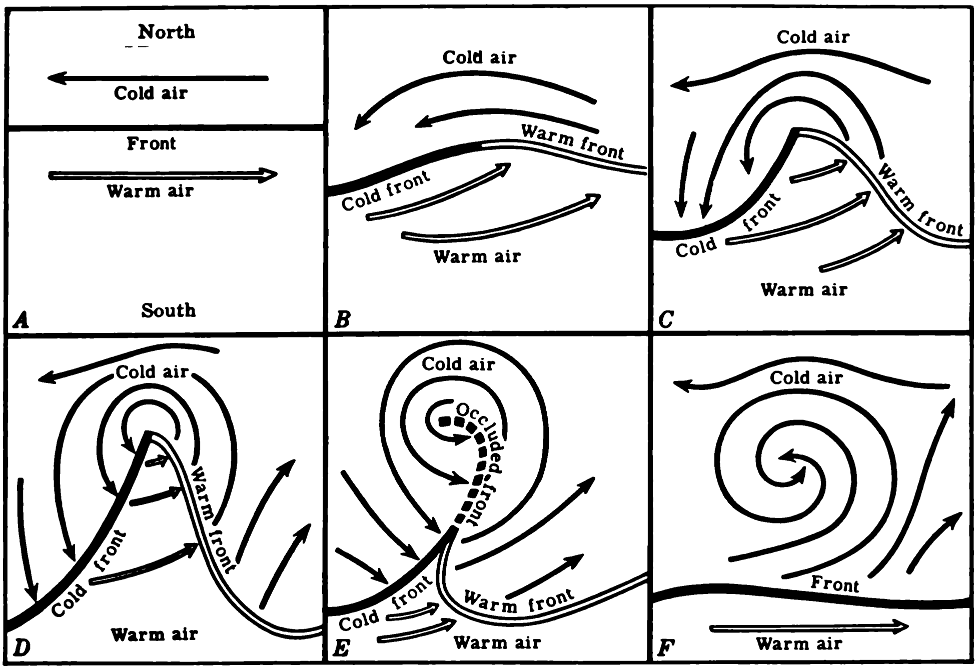


Figure 5.6
Plan view of stages in the development and occlusion of an extratropical cyclone along the polar front in the Northern Hemisphere. A, *initial stage;* B, *beginning of cyclonic circulation;* C, *warm sector well defined between fronts;* D, *cold front overtaking warm front;* E, *occlusion;* F, *dissipation.* (NOAA)

THE EXTRATROPICAL CYCLONE

The general circulation of the atmosphere which favors divergence and air mass development in tropical and polar latitudes is characterized in the mid-latitudes by convergence of dissimilar air masses. When air masses having different temperature and humidity properties come together they do not mix readily but maintain a boundary surface of discontinuity for some time, the warmer, lighter air being forced aloft over the colder mass. The sloping boundary surfaces between contrasting air masses are called *fronts*. On the weather map a front is indicated by a line which represents the intersection of the frontal surface with the ground, but it should be remembered that a front is three-dimensional. It extends vertically as well as horizontally and it has thickness. Frontal boundaries vary from 2 or 3 to more than 50 km in width. In the zone of contact between the prevailing westerlies and the polar easterlies there is a more or less permanent, undulating frontal discontinuity known as the *polar front*. Along this polar front the extratropical (mid-latitude) cyclones are developed. In much the same way that whirlpools are formed between adjacent currents of water moving at different speeds, giant waves or whirls form along the polar front (see Figure 5.6).

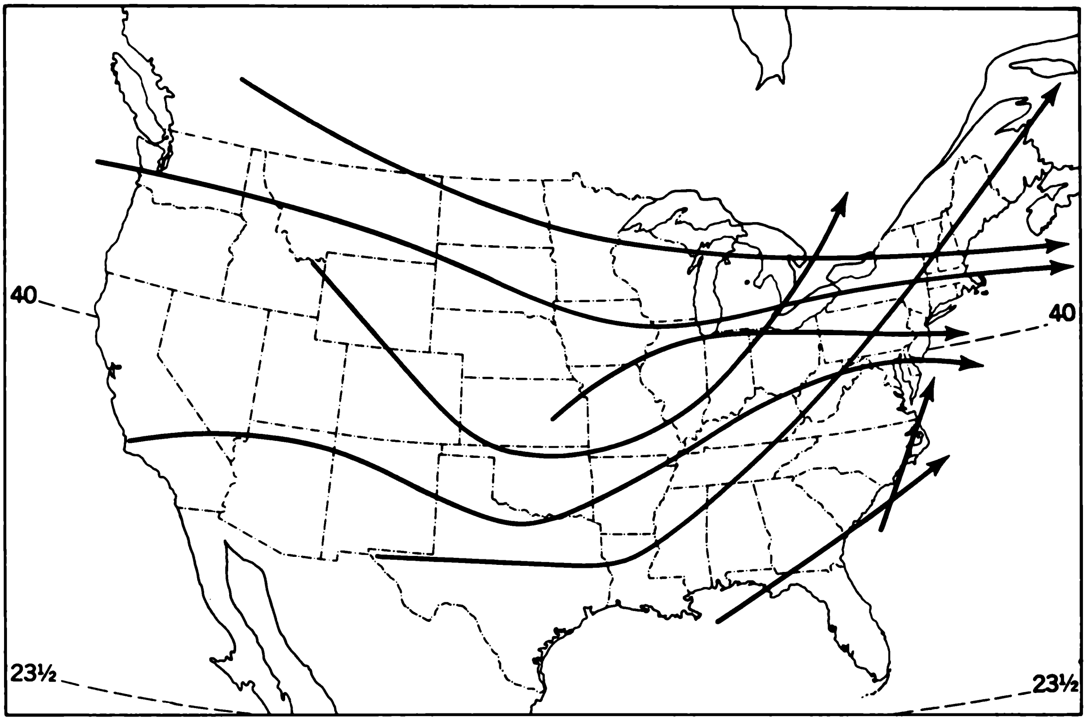


Figure 5.7
Common paths of extratropical cyclones across the contiguous United States.

In the typical Northern Hemisphere cyclone the cold air mass pushes southward along the ground under the warm air, which advances northward. The resulting convergence and rising air along the front is accompanied by low pressure, and it is for this reason that cyclones are often called lows or depressions. As the cyclonic wave develops, the pressure gradient is focused toward the center and the pattern of air flow takes on a counterclockwise, or cyclonic, circulation with winds making angles of 20° to 40° with the isobars. Note that the Coriolis effect deflects the winds to the right of their intended path as they blow toward the low center. (In the Southern Hemisphere the Coriolis effect produces a clockwise circulation into the low center.) The "tongue" of warm air advancing from the south is known as the *warm sector.* Where the advancing warm air mass is replacing colder air at the surface the boundary is called the *warm front.* To the west of the warm sector the leading edge of the cold air is the *cold front.* Because air is being forced to rise along these fronts they are accompanied by appreciable cloudiness and usually precipitation. Conditions along the fronts are spoken of as "frontal weather," whereas "air mass weather" prevails in the areas away from the fronts.

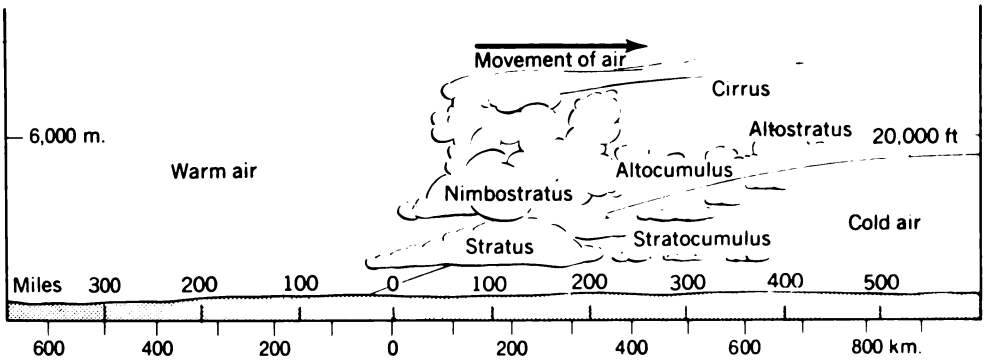
Extratropical cyclones vary in diameter from 200 to as much as

3,000 km. Most are in the range of 800 to 1,500 km. Their form may be roughly circular or elongated and oval, or they may be broad, shallow, weak depressions. Usually they travel in groups, or "families," whose movements follow waves in the pressure pattern at upper levels. Their general direction of movement is from west to east in the mid-latitude westerlies, but specific paths are often curved and sometimes erratic. Over the North American continent they tend to curve toward the southeast into the Mississippi Valley and then toward the northeast (Figure 5.7). The speed of an extratropical cyclone also varies; the average is about 30 to 50 km per hr or from 800 to 1,100 km per day. The rate of movement is greater in winter than in summer. As long as the discontinuity in temperature and moisture conditions is maintained along its fronts the cyclone is likely to persist, but a cyclonic pattern of pressure and winds is also essential. Some cyclones pass from central North America out across the Atlantic and enter western Europe; others dissolve because of modifications of their energy and moisture properties.

THE WARM FRONT

The warm front normally extends east and southeast from the low center of a cyclone, in the Northern Hemisphere. The wind shift at the warm front is not so pronounced as at the cold front. Maximum wind speed, perhaps with gustiness, is experienced slightly in advance of the front. Barometric pressure decreases gradually until the front passes and then tends to level off. The slope of the front is gradual; 300 km ahead of its intersection with the ground it may be only 1,500 or 2,000 m above the surface. The great areal expanse of rising air produces a vast cloud system with the highest clouds lying far in advance of the front. Since the vertical movement along most of the warm front boundary is not violent the types of clouds are predominantly stratiform. High wispy cirrus, often in the form of "mares' tails" appear first. As the front moves in, the clouds lower and thicken progressively to cirrostratus, altostratus, and nimbostratus. (See Figure 5.8).

Figure 5.8
Vertical cross section of a warm front. (NOAA)



Precipitation from the warm front is light to moderately heavy and continuous, extending over a wide zone. Fogginess and poor visibility are common in the precipitation area. Sometimes the typical warm front precipitation is interspersed with heavy showers where convection takes place in unstable warm air as it rises rapidly over the cold air just ahead of the front. In winter, ice pellets or freezing rain may occur when rain falls from the warm air through the cold mass below. The succession of weather can include snow, ice pellets, and then rain as the front approaches, clouds lower, and temperature rises. Sometimes the temperature does not rise above freezing during the entire passage of a warm front. Nevertheless, an appreciable rise in surface temperature normally does accompany the passage of the front and the relatively warmer, moister air of the warm sector then begins to dominate the weather. The air which occupies the warm sector is likely to be unstable and therefore produce convective showers, especially in summer over land that is being heated.

THE COLD FRONT

The cold front is the leading edge of an intrusion of cold air into territory previously occupied by warmer air. It pushes under the warmer air after the fashion of a wedge, forcing it to rise. Frictional drag along the ground retards the advance of the cold air and it develops a relatively steep forward surface in contrast to the gentler slope of the warm front. (See Figure 5.9.) Hence it is accompanied by clouds with vertical development and heavy, showery precipitation. At times, cold air in the upper levels overruns the warm air to create extreme instability and overturning. Convective clouds and showers then precede the surface advance of the front.

The most significant identifying feature associated with the cold

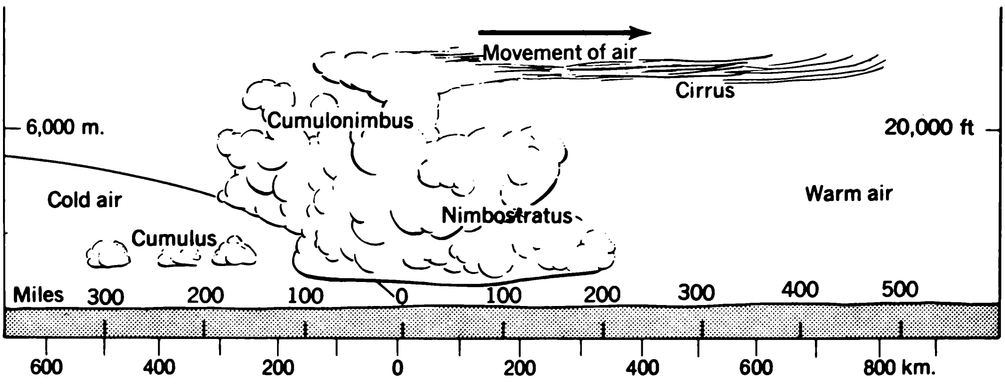


Figure 5.9
Vertical cross section of a cold front. (NOAA)

front is the wind shift at its passage. In the warm sector ahead of the front, winds are generally southerly or southwesterly. As the front passes, the wind shifts to the northwest or north. The approach of the cold front is heralded by falling pressure; often there is a sharp drop just as the front passes. This is the basis for the popular association of a falling barometer with a coming storm. As the front moves on, the barometer rises again. A drop in temperature can ordinarily be expected with the passage of a cold front, but the surface temperatures are not so reliable as those in the upper air for detecting the transition from warm to cold air. Because of the lower temperatures and the probable polar origin of the air behind the cold front, it will have a lower moisture content. Therefore, humidity values also help in location of the front. The zone of precipitation along the cold front is ordinarily much wider than the discontinuities of wind, temperature, and humidity would seem to indicate, and yet it is narrower than in the warm front. A few hours after the front has passed, clearing weather can normally be expected. The following cold air mass moves over warmer ground which is probably also moist, so that instability may develop in its lower levels with resulting scattered showers. If the cold mass is unstable and moist itself, and especially if it is moving over mountainous terrain, intermittent cloudiness and showers may continue for several hours or for a day or two. In winter, when the precipitation is snow, the combination of low temperatures, high winds, and blowing snow along and following the cold front is called a *blizzard* in the central United States. A similar storm in Russia is known as the *buran*.

OCCLUDED FRONTS

In the cyclonic storm, the cold front usually advances faster than the warm front, eventually overtaking it. When the air mass behind the cold

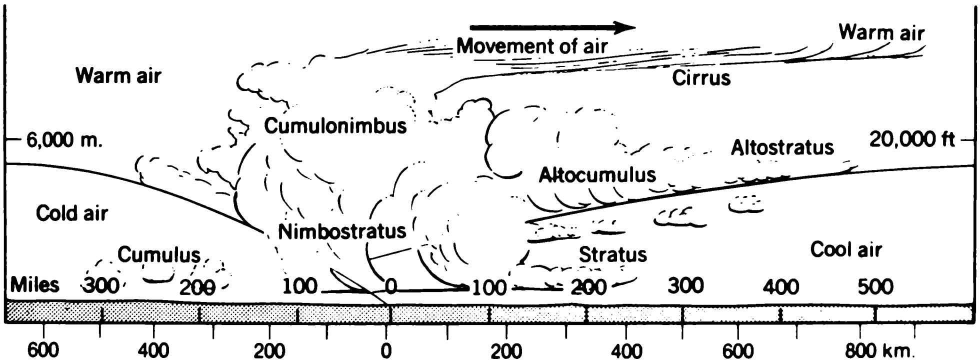


Figure 5.10
Vertical cross section of a cold-front occlusion. (NOAA)

front comes into contact with the air in advance of the warm front, the air of the warm sector, being less dense, is pushed aloft. This process is called *occlusion,* and the new frontal surface thus formed is an *occluded front.* There are two ways in which an occluded front may form, depending upon the relative temperatures of the air masses within the cyclone. Over continents and east coasts of continents the air behind the cold front is normally colder than that ahead of the warm front because it has had a shorter path across relatively warm surfaces. As the occlusion develops, the cold front will run under the cool air to form a *cold-front occlusion* (Figure 5.10). Prefrontal weather ahead of the cold-front occlusion is similar to that of the warm front. The lifting of conditionally unstable air in the warm sector may be the "trigger action" to produce thunderstorms. At the front there is typical cold-front activity, and after the occlusion has passed a marked improvement in conditions occurs. In their later stages, cold-front occlusions often become well-defined cold fronts. This results from the introduction of increasingly colder air behind the front and increasingly warmer air in the "cool" area ahead of the front.

Where the air behind the cold front is warmer than that ahead of the warm front, the former (cool) will run up over the latter (cold) when occlusion occurs, producing a *warm-front occlusion* (Figure 5.11). In winter along west coasts where cool air flowing onshore is usually warmer than the cold air over the land, warm-front occlusions are especially prevalent. In its initial stage, this type of occlusion shows most of the characteristics of the warm front. Then follows moderate frontal activity in connection with the *upper cold front.* As the occlusion process continues, the whole system tends to dissipate because there is a lack of sharp, persistent differences in air mass properties.

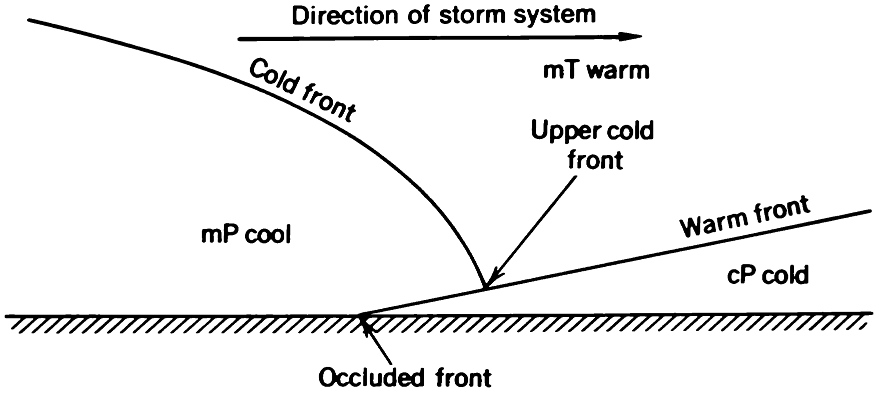


Figure 5.11
Vertical cross section of a warm-front occlusion.

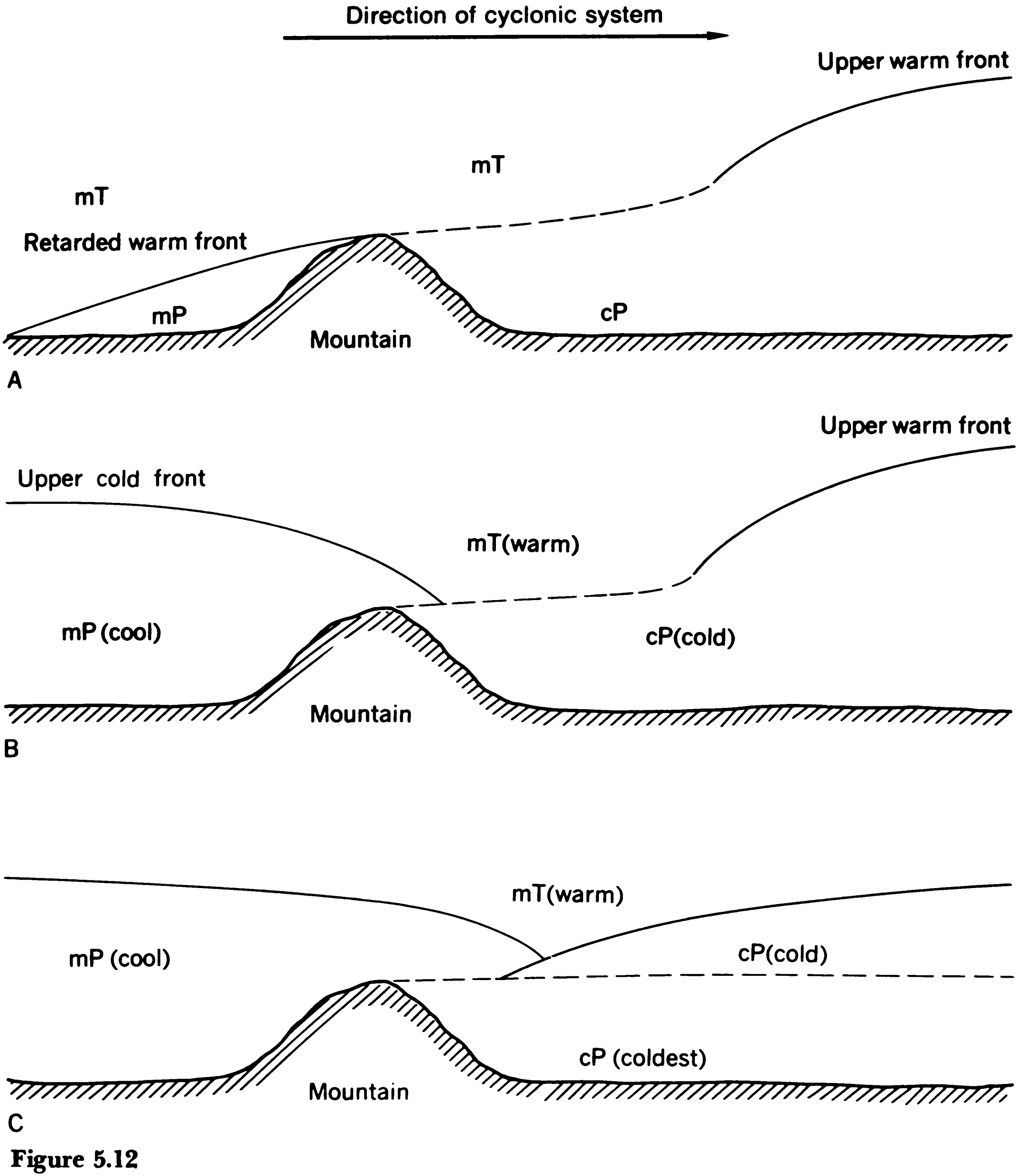


Figure 5.12
Stages in occlusion and formation of upper fronts induced by a mountain barrier. (After Willett)

MOUNTAIN BARRIERS AND SURFACE FRONTS

Mountain barriers lying across the path of a moving cyclone tend to induce occlusion by delaying the movement of the cold air in advance of the warm front and by promoting more rapid lifting of the air in the warm sector. The tendency to occlude depends upon the relative stability of the air masses involved. Stable air is dammed more effectively than unstable air, which may become even more unstable when forced up mountain slopes. This phenomenon is common on the west coast of North America, where numerous cyclones occlude as they move against the mountain ranges. When, as is frequently the case, a warm-front occlusion is formed, the moist, unstable air off the ocean moves over the mountain barrier and may continue for some distance to the leeward as an upper cold front, riding above the cold continental air mass which preceded the original warm front. (See Figure 5.12.)

In addition to retarding the advance of frontal systems, mountains produce greater precipitation on their windward slopes than on the leeward as a result of the orographic effect.

ANTICYCLONES

The term *anticyclone* implies characteristics opposite those of the cyclone. Barometric pressure is highest at its center and decreases outward. Consequently the anticyclonic wind system blows out from the center, and because of the Coriolis effect it has a clockwise circulation in the Northern Hemisphere (counterclockwise in the Southern Hemisphere).

Further, the anticyclone is composed of subsiding air which renders it stable in contrast to cyclones or other low-pressure systems. It rarely has fronts or definite wind-shift lines, but a gradual change in wind direction takes place as it passes. Except where surface instability is induced (for example, as the air moves over a warm surface) there is less cloudiness and therefore lack of precipitation in the typical anticyclone. Wind shear, convergence, and frontal activity often develop in the trough of low pressure between two anticyclones, however (Figure 5.13).

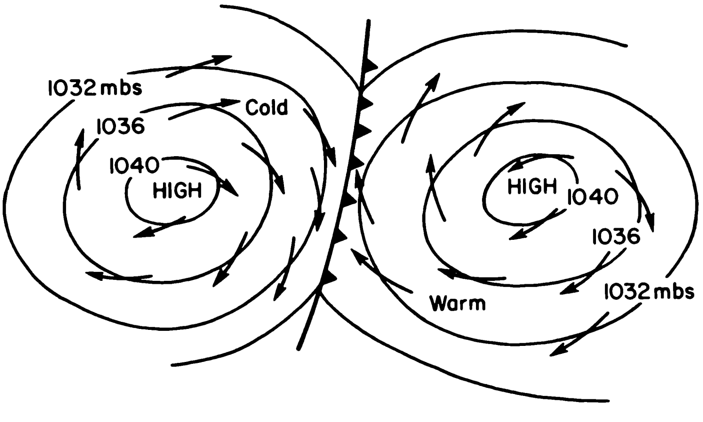


Figure 5.13
Cold front in a trough of low pressure between two adjacent anticyclones, Northern Hemisphere.

In winter in middle and high latitudes anticyclones are essentially synonymous with the source regions of cold stable air masses which invade cyclonic systems in the form of cold fronts. These are known as "cold" anticyclones and they are confined to the lower troposphere. Like the cyclone, the anticyclone usually has a pressure pattern represented by circular or oval isobars, but it may assume various shapes. Diameters range from a few hundred to 3,000 km. On the average it travels at a rate appreciably lower than the typical cyclone, but it is even more erratic than the cyclone in its direction. The paths of highs are roughly similar to those of lows across North America except that they do not usually turn northeastward near the Atlantic Coast but proceed more directly out over the ocean.

The anticyclones of the subtropics are associated with the subtropic highs in the general circulation and are warmer than those of the higher latitudes. Their pressure results from the piling up of a great mass of air; subsidence accounts in part for their warmth. Although they tend to be more nearly stationary, portions sometimes break away to move along the margins of the westerlies. They are frequently reinforced by polar anticyclones that merge with them. Because of the difference in temperature between the two, a front may be formed in the trough of lower pressure that separates the high cells before they completely merge. Along their trade-wind margins waves may develop in the pressure pattern, leading to cloudiness and precipitation.

TROPICAL STORMS

Although they have certain features similar to those of mid-latitude storms, tropical disturbances generally do not exhibit sharp discontinuities of temperature. Many have weak pressure gradients and lack well-defined wind systems. Extensive, shallow lows occasionally bring long periods of overcast weather with continuous rain. In the intertropical convergence there may be convective activity and thunderstorms. Convergence tends to increase when the equatorial trough of low pressure moves poleward in summer, producing bands of cumulonimbus clouds and high overcasts of cirrus.

A common feature of tropical weather is the *easterly wave,* which normally forms in the trade winds and moves slowly from east to west (see Figure 5.14). Fair weather and divergent winds precede the wave, followed by convergent air flow and an extensive belt of towering clouds that may appear similar to a mid-latitude cold front. Squally weather and precipitation frequently accompany such a disturbance. Some easterly waves move poleward and curve toward the east to become extra-

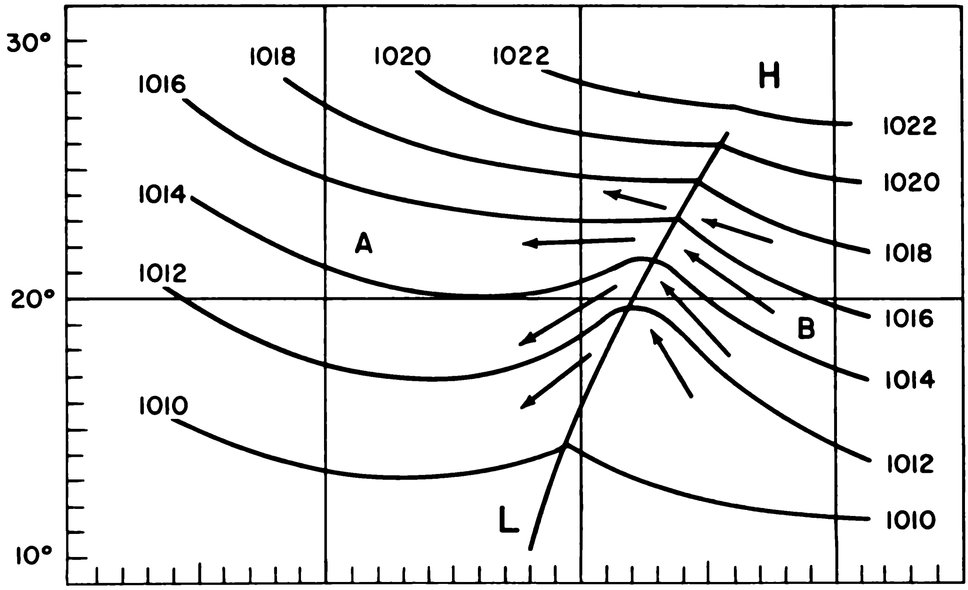


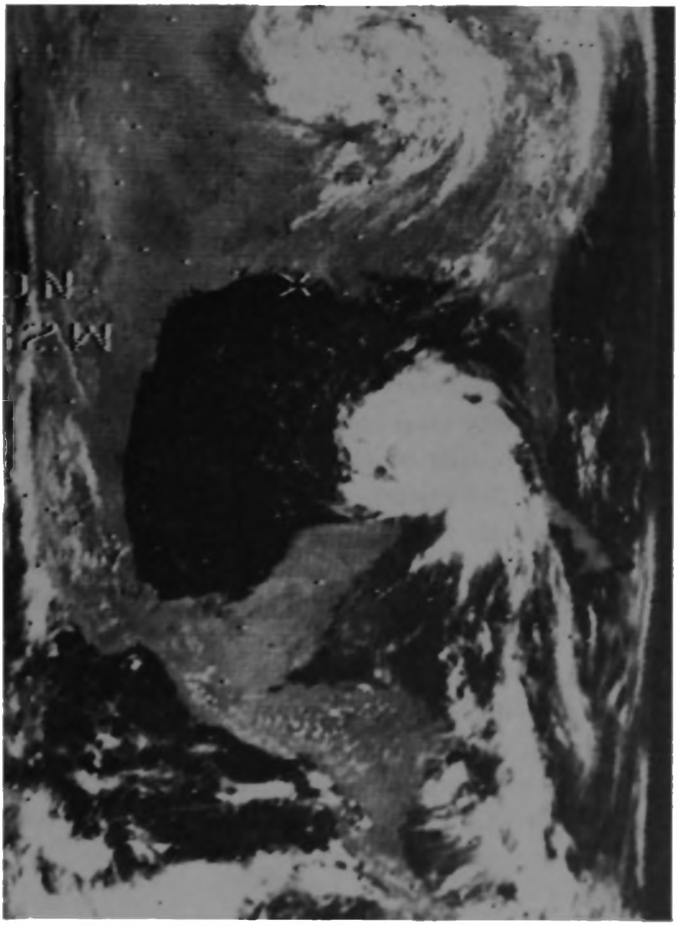
Figure 5.14
Pressure distribution in an easterly wave in the Northern Hemisphere. Winds blowing from the direction of B converge toward the wave; they diverge in the region of A.

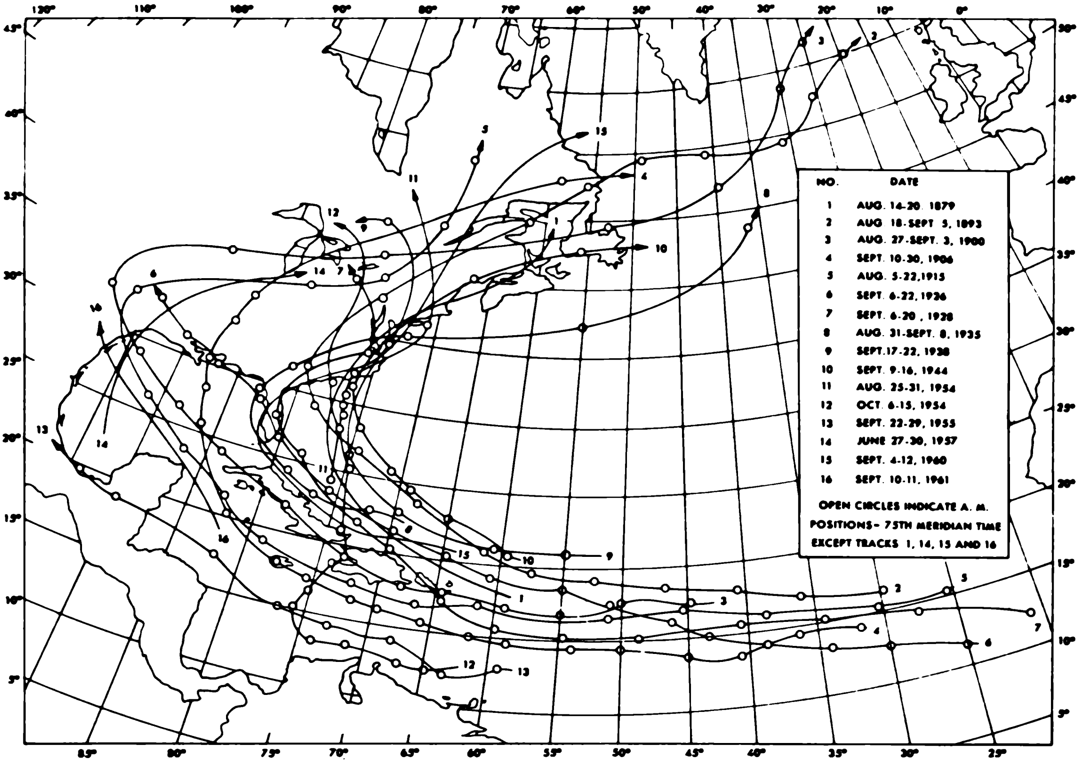
tropical cyclones. Others may develop vortices, become tropical cyclones, and even grow to hurricane intensity.

The violent and destructive forms of tropical cyclones are much better known than the weaker variety although the former are, fortunately, much less common. They originate over the tropical oceans only. In the Caribbean and off the Pacific Coast of Mexico they are known as *hurricanes* (Figure 5.15); in the seas off China, the Philippines, Japan, and the other islands of the western Pacific they are called *typhoons;* in the Indian Ocean they are simply called *cyclones,* a term which should not be confused with cyclones in general. In the Southern Hemisphere they occur east of the African coast and along the northwest and northeast coasts of Australia. Off northwest Australia the associated strong winds are locally known as *willy-willies.* Elsewhere the term *tropical cyclone* is generally applied. To avoid possible confusion with weak tropical cyclones and the extratropical cyclones, the term *hurricane* will be used in the following paragraphs unless otherwise specified. Hurricanes are apparently absent from the South Atlantic, presumably because the equatorial belt of convergence seldom moves far enough south of the equator in that region.

A hurricane is a giant heat engine that derives its energy mainly through the transfer of sensible and latent heat from sea to air. It takes its whirling motion from the Coriolis effect and consequently is most likely to form in the zone of intertropical convergence when the latter is located at least 5° to 10° from the geographic equator, that is, in late summer or early autumn. At the equator the deflective force is inadequate to generate the violent vortex. Most hurricanes evolve from preexisting disturbances such as incursions of cold air in the upper troposphere, easterly waves, or weak tropical depressions that become charged with warm, moist air and migrate poleward. Off the east coasts of Asia and North America they often move northwestward and then turn away toward the northeast, although the actual paths vary widely and may be erratic (see Figure 5.16). The diameter of the typical hurricane is 150 to 1,000 km, and it increases as the storm moves away from the low latitudes. Some hurricanes, however, have diameters as small as 30 km. The size of the disturbance has no direct relation to intensity, and its rate of travel appears to be unrelated to either its size or the wind speeds within the system. An advance of 15 to 30 km in an hour is typical, but hurricanes are erratic in their general rate of progress as well as in direction.

Figure 5.15
NIMBUS III photograph of Hurricane Camille *entering the Gulf of Mexico, August 16, 1969.*
Clouds of a mid-latitude depression lie over the Mississippi Valley. Hurricane Camille *subsequently passed inland over the coasts of Louisiana and Mississippi and northward as far as Kentucky before turning eastward.* (NASA, Goddard Space Flight Center)





NO.	DATE
1	AUG. 14-20. 1879
2	AUG 18-SEPT 5, 1893
3	AUG. 27-SEPT. 3, 1900
4	SEPT. 10-30, 1906
5	AUG. 5-22,1915
6	SEPT. 6-22, 1926
7	SEPT. 6-20 , 1928
8	AUG. 31-SEPT. 8, 1935
9	SEPT.17-22, 1938
10	SEPT. 9-16, 1944
11	AUG. 25-31, 1954
12	OCT. 6-15, 1954
13	SEPT. 22-29, 1955
14	JUNE 27-30, 1957
15	SEPT. 4-12, 1960
16	SEPT. 10-11, 1961

OPEN CIRCLES INDICATE A. M. POSITIONS- 75TH MERIDIAN TIME EXCEPT TRACKS 1, 14, 15 AND 16

Figure 5.16
Paths of some destructive hurricanes in the North Atlantic. (Department of the Army)

Several features distinguish tropical hurricanes from the cyclones of the mid-latitudes. The pressure distribution, which is represented by isobars, is more nearly concentric and circular and the isobars are closely spaced with very low values at the center, indicating the steep pressure gradient which produces the high-velocity winds. There are no fronts or wind-shift lines, but at the center of the whirl there may be a calm "eye" 10 to 50 km in diameter in which air is descending and which, therefore, is comparatively clear and warm. The lack of introduction of contrasting air masses results in fairly even distribution of temperature in all directions from the center. Rainfall is also relatively evenly distributed, especially if the storm is stationary; in a moving hurricane, rainfall is slightly greater in the forward half of the storm. In either case rain is torrential. Because of the tropical nature of the air and the high freezing level in the latitudes of hurricanes, hail does not occur.

The outstanding feature of the tropical hurricane is its wind force. At the outer margins of the system winds are only moderate, but velocities increase rapidly toward the center, reaching a maximum along the outer edge of the central eye. Technically, wind speed must reach 65 knots (33 m per sec or 75 mi per hr) to be classified as of hurricane force;

speeds of 130 knots have been recorded, with gusts estimated as high as 220 knots.

The destructiveness of hurricanes is due to extreme wind force, which crushes frame structures, blows over standing objects, and sends huge pieces of debris flying through the air. Heavy rainfall and flooding also cause extensive damage. At sea, hurricanes produce a distinctive heavy swell that affects ocean shipping. Swells in advance of an approaching hurricane come at longer intervals than ordinary storm swells and are one of the most valuable warning signs to mariners because they appear before the winds of the hurricane are encountered. Heavy waves accompanying the hurricane swell can cause coastal damage even while the storm center is well out to sea. As the storm approaches a coast, the piling up of water by strong winds may produce a disastrous *storm surge.*

When hurricanes pass inland, frictional effects and the loss of sustaining energy from warm water surfaces cause them to lose much of their force. Although of interest to the meteorologist, this phenomenon may be a small consolation to people in the affected region. As it moves to higher latitudes, where prolonged rain may produce flooding long after winds have moderated, a hurricane often draws in air from poleward sources and transforms into an extratropical cyclone.

As a matter of convenience for tracking the paths of hurricanes in the western Atlantic and typhoons in the Pacific, the weather services name the storms in alphabetical order as they are discovered.

THUNDERSTORMS

Several thousand thunderstorms occur every day, mainly in the tropics. In polar regions they are virtually unknown. They are always associated with unstable air and strong vertical motion that produce clouds of the cumulonimbus type. A great deal of the energy for their development comes from the release of the latent heat of condensation in rising humid air. It is important to note in this connection that an unstable lapse rate and overturning of air may result from either warming of surface layers or introduction of cold air aloft.

The common processes that initiate thunderstorm development are (1) heating and convection in moist air over warm land surfaces; (2) passage of cold, moist air over warm water; (3) forced ascent of conditionally unstable air along zones of convergence or at mountain barriers; or (4) radiational cooling at upper levels. Moisture and rising air are the essentials for any thunderstorm. Most thunderstorms consist of several convective cells that grow and dissipate erratically. When the

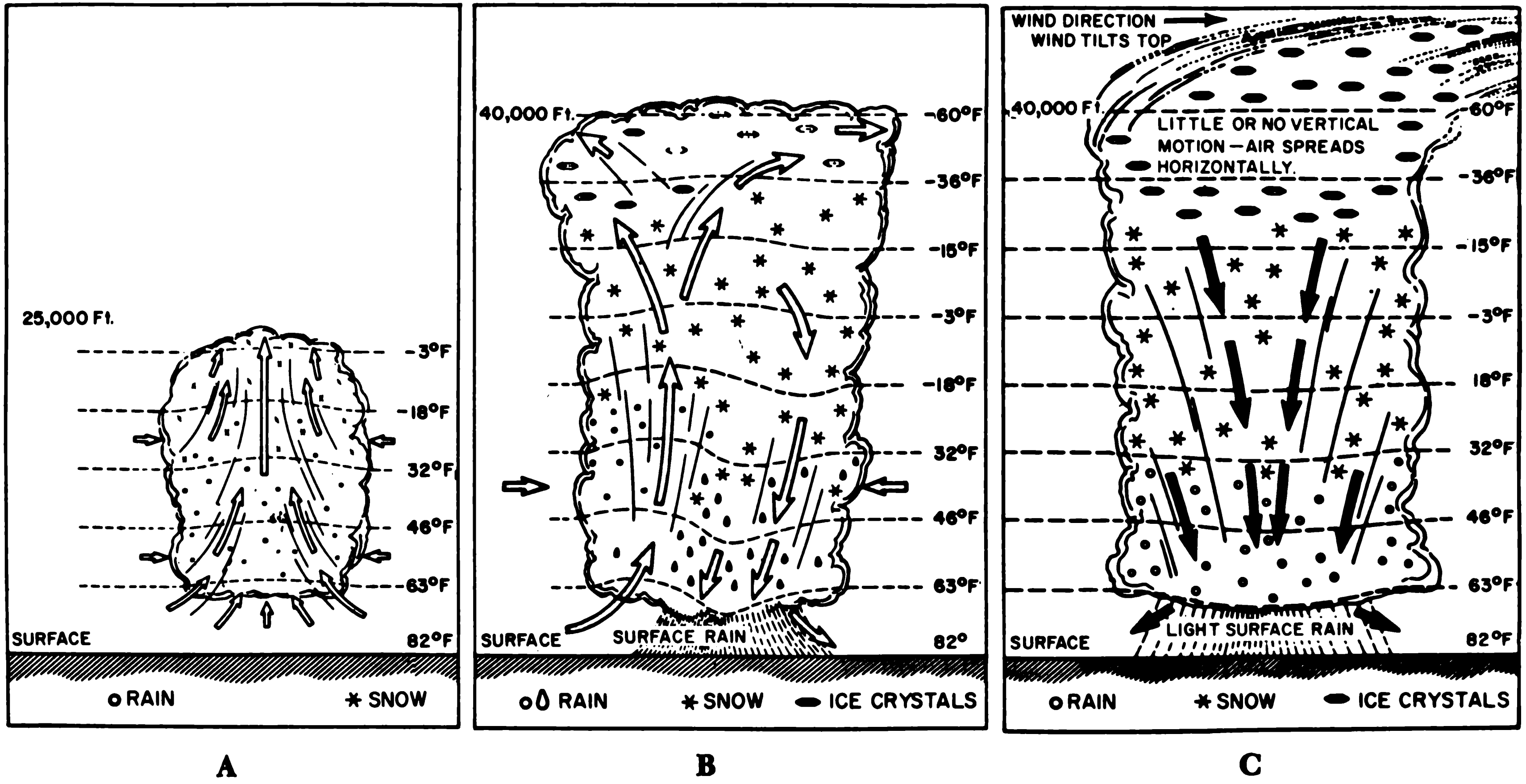
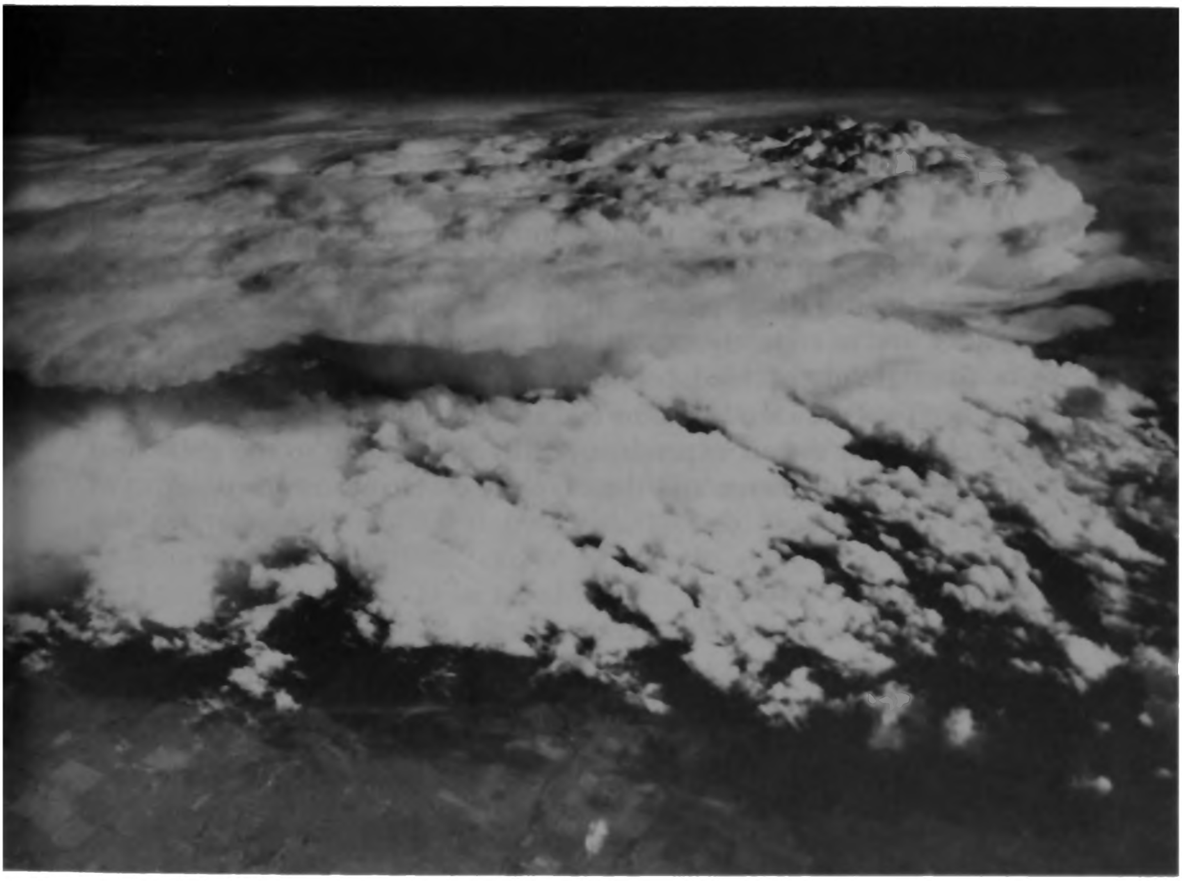


Figure 5.17
Evolution of a thunderstorm. A, *early (cumulus) stage;* B, *mature stage;* C, *dissipating stage.*

force of rising air is great enough the storm can reach to heights from 4 to more than 20 km (see Figures 5.17 and 5.18). Cooling proceeds at the dry adiabatic rate up to the base of the storm cloud and continues to be rapid as dry air is entrained in the violent updrafts. Inertia in the convective "chimney" may carry air beyond the level where equilibrium is established between the convective mass and the surrounding air aloft. The height of a thunderstorm is related to latitude and season of the year; the greatest heights are developed in summer and in the tropics. Typical diameters of thunderstorms vary from 3 to 40 km. A well-developed cumulonimbus thunderhead has an anvil-shaped crest that points in the direction of the storm's movement, and there are often shelves of clouds in advance of the main thunderhead. Along its base the storm is dark and ominous, and it may be preceded by roll clouds of a *squall line,* created by air currents moving in opposing directions. Precipitation from the mature storm is intense and composed of large raindrops, literally a cloudburst. If the updrafts have sufficient force and

Figure 5.18
Aerial view of a mature thunderstorm. The top of the anvil is about 15 km above the ground. Note the convective humps on the upper surface of the cumulonimbus and the shelf clouds extending toward the lower right. (NOAA, National Severe Storms Laboratory)



penetrate well above the freezing level, hail may fall from the cloud, usually from the leading edge. Under favorable temperature conditions precipitation may be in the form of snow or snow pellets.

Thunderstorms caused by surface heating over land are most common in summer and in the afternoon or early evening. Over the oceans the temperature difference between the water and the cooler air above is greatest at night, and thunderstorm activity resulting from this set of conditions is accordingly greater at night. Along and above mountains maximum incidence of thunderstorms is usually in the afternoon or early evening, when the combined effects of daytime heating and orographic lifting are at a maximum.

Along zones of convergence thunderstorms develop when air is forced up rapidly. In the mid-latitudes they may be associated with the cold front, warm front, or upper fronts and are often accentuated by surface heating, orographic lifting, or overrunning by cold air at upper levels. Along a cold front they are generally closer to the ground and more violent than the warm or upper-front types. In contrast with localized convective thunderstorms, which tend to be spotted about in an erratic pattern, the frontal types are concentrated in a zone 20 to 80 km in width and perhaps hundreds of kilometers in length. Because the upper portions of cumulonimbus clouds are frequently obscured by lower clouds, frontal thunderstorms are sometimes difficult to identify from the ground. Pilots flying at high altitudes have some advantage in this respect. A line of air mass thunderstorms along the crest of a mountain range should not be confused with frontal thunderstorms, although it does indicate convergence of air above ridges and peaks.

For the purposes of meteorological records, thunder must be heard or overhead lightning or hail observed at a station before a thunderstorm is reported. Although thunder and lightning accompany the typical mature thunderstorm and often are its most dramatic manifestations, their roles in the development of precipitation and other storm characteristics are not clearly understood. Lightning discharges may take place from cloud to cloud horizontally, between different levels in a cloud, or from the cloud base to the ground. Thunder is the explosive sound created as the air expands suddenly in response to the great heat of the lightning discharge and then rapidly cools and contracts.

TORNADOES AND WATERSPOUTS

Tornadoes are the most violent storms of the lower troposphere. A tornado is a tight vortex, gyrating around a center of extremely low pressure. Accurate wind observations are impossible within the mael-

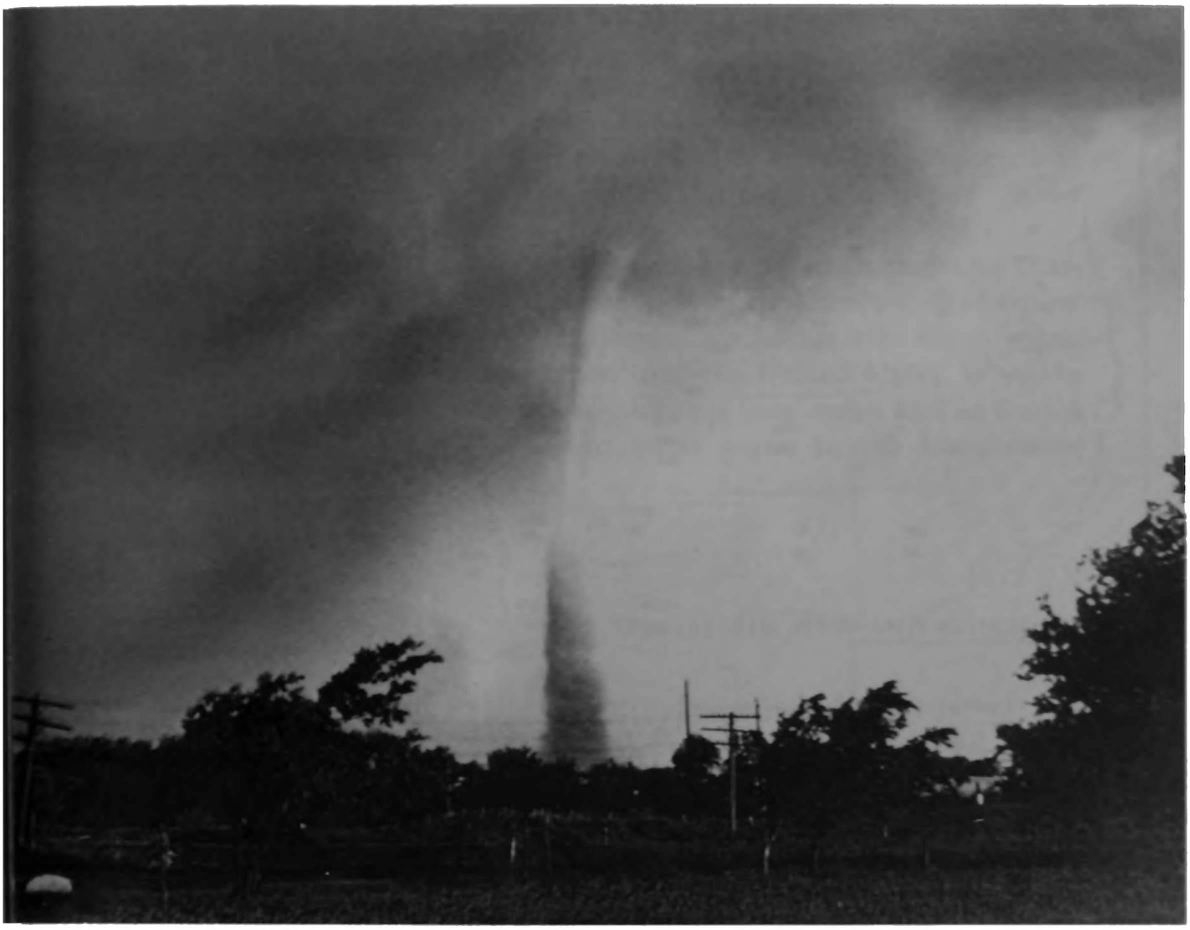
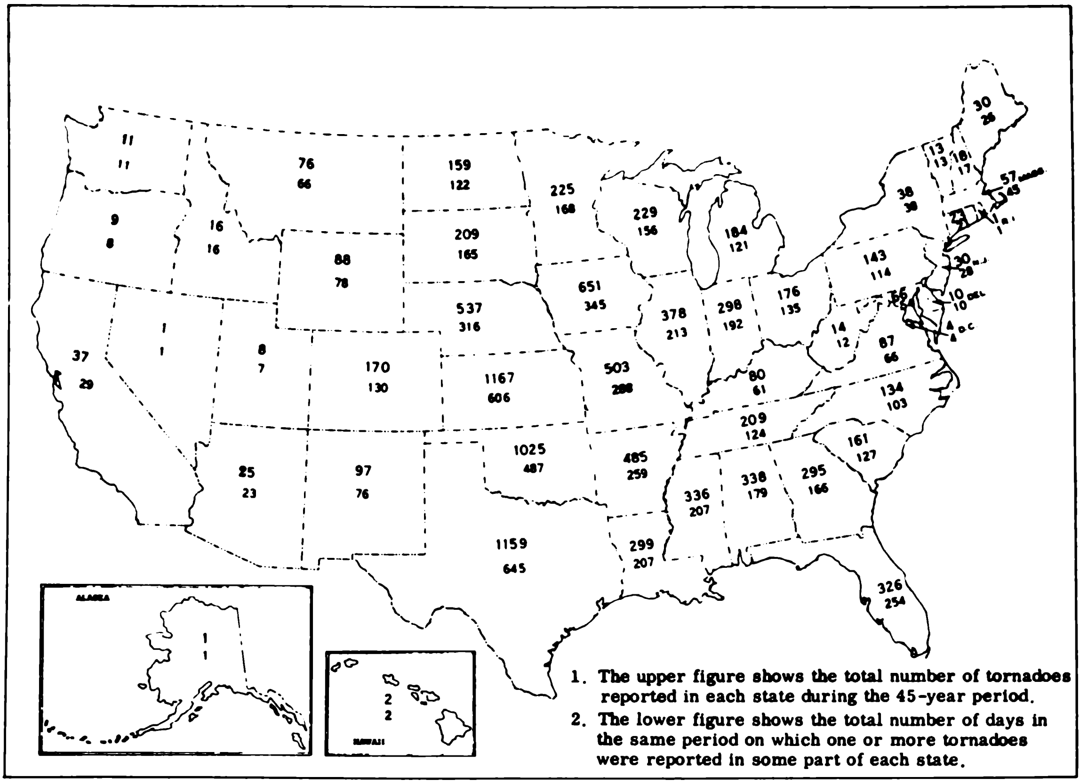


Figure 5.19
Tornado near Enid, Oklahoma, June 5, 1966. The upper funnel consists of condensation particles; the lower column is dust and debris raised by the whirling vortex. (Photograph by Leo Ainsworth, courtesy NOAA, National Severe Storms Laboratory)

strom, but velocities of 200 to 300 knots are probably common. Estimates of windspeed are based on surveys of damage and analyses of motion picture records. A tornado can be distinguished by its writhing, funnel-shaped cloud, which extends downward from the base of a cumulonimbus or a turbulent cloud layer (Figure 5.19). If it reaches the ground incredible destruction is suffered by anything in its path. Buildings seem to explode as a result of the wind force and the sudden decrease of outside pressure. A tornado that travelled through Missouri, Illinois, and Indiana on March 18, 1925, caused 689 deaths and property losses estimated at $16.5 million. Most tornadoes are only a few hundred meters in diameter at the ground, but observers have recorded swaths 3 to 5 km wide. In mid-latitudes the funnels form in the warm sector of cyclonic systems just ahead of the cold front where dry air is overrunning aloft and sometimes along the leading edges of intense thunderstorms. Tornadoes also occur in hurricanes; an astounding total of 115 were



1. The upper figure shows the total number of tornadoes reported in each state during the 45-year period.
2. The lower figure shows the total number of days in the same period on which one or more tornadoes were reported in some part of each state.

Figure 5.20
Distribution of reported tornadoes in the United States, 1916–1960. Note that the values are totals for states, which vary greatly in area. (NOAA, National Environmental Data Service)

observed in the company of Hurricane *Beulah* over southeastern Texas in September, 1967. Thunderstorm activity, rain or hail, and lightning accompany most tornadoes. Low-level convergence and abnormally strong convective turbulence in moist, unstable air presumably generate the vortex. Paths of tornadoes usually are parallel to the cold front and toward the center of low pressure in an extratropical cyclone. This rule has exceptions, however, for they have been observed to make U-turns and even complete circles. Their rate of travel at the ground may reach 50 knots or more or they may remain stationary for short periods. Some make contact with the ground and then lift, only to strike again at a distance of several kilometers; others travel only a few meters before rising and dissipating.

In North America the greatest frequency of tornadoes is in the Mississippi Valley and Great Plains, but they also occur in the Gulf States, western Lake States, and the Canadian Prairie Provinces (Figure 5.20). One was sighted at Yellowknife, N.W.T., Canada (lat. 62°28′ N.) on June 19, 1962. Every state in the contiguous United States has experienced tornadoes. They are most likely to appear in spring and early summer, when contrasts of temperature and humidity in the air masses along cold fronts are at a maximum. Although tornadoes are apparently less common in other parts of the world, they have been reported at widely scattered locations in the middle latitudes and tropics. Many undoubtedly have escaped official climatic records.

At sea tornadoes become *waterspouts,* having much the same characteristics except that they are usually smaller in diameter. Waterspout frequency is greatest where cold continental air pushes over warm water, as off the east coasts of China, Japan, and the United States. When in contact with the surface a waterspout picks up some spray, but its funnel is composed primarily of condensed water vapor in the low-pressure vortex.

SPECIAL AIR MASS AND STORM EFFECTS

There are a number of weather phenomena which do not possess the dramatic features generally considered to belong to storms and which are not accompanied directly by precipitation. Yet they incorporate distinctive "weather," and a few are referred to as storms, at least locally.

Perhaps the best known of these phenomena is the *foehn effect,* which produces a warm, dry, and often gusty wind on the lee side of mountain ranges. The name *foehn* originated in Austria and Germany, where the foehn wind is frequently experienced in valleys of the Alps. In the western United States and Canada the same type of mountain-induced effect is called the *chinook.* The explanation of the relative warmth of the typical chinook rests upon two principles. Ordinarily a chinook wind is accompanied by cyclonic activity which produces clouds and precipitation on the windward side of the mountain range (for example, the Rockies). The latent heat released to the air by the condensation process warms the air which passes across the range, and, because the air has lost some of its moisture, it will also be drier (see Figure 5.21). However, the latent heat of condensation alone does not account for the temperatures which occur on the leeward. The mountain barrier creates a frictional drag which tends to pull the air from higher levels down on the leeward. Air forced down in this way is heated adiabatically, and therefore its relative humidity is lowered. This action is by no means regular, especially in its earlier stages, but comes in surges which are experienced at the ground as gustiness. If the wind is to affect an extensive area to the leeward of the mountains the general pressure gradient must be such that the cold air will gradually move out ahead of the chinook. When fully developed the chinook can remove snow cover in a short time, and it is in winter that it is most often recog-

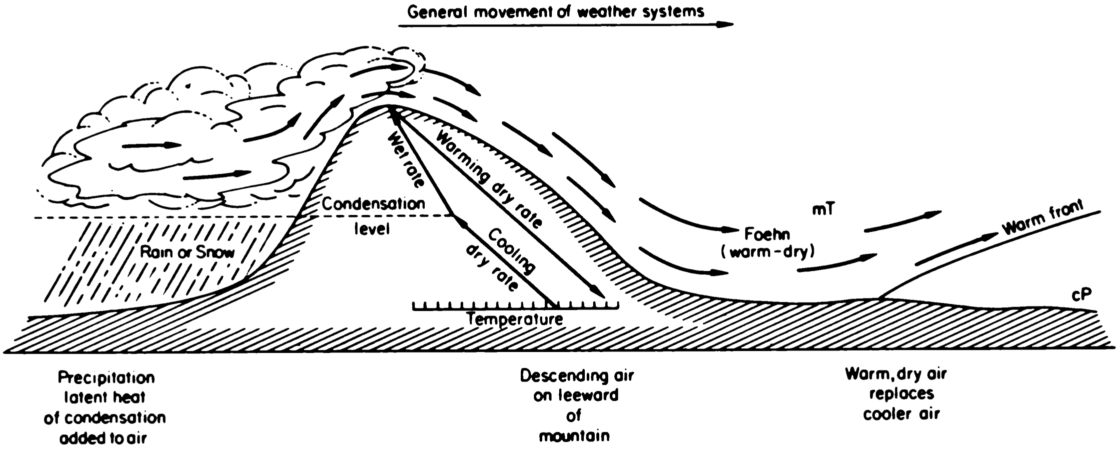
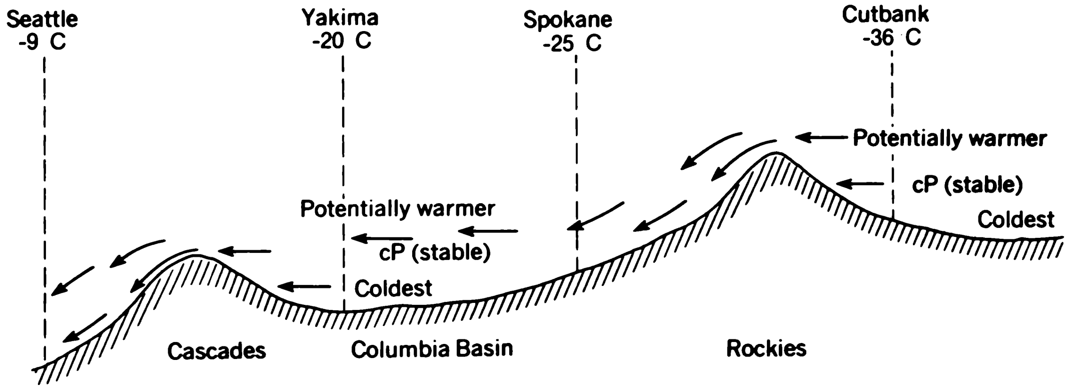

General movement of weather systems
Wet rate
Warming dry rate
Cooling dry rate
Condensation level
Rain or Snow
Temperature
Foehn (warm-dry)
mT
Warm front
cP
Precipitation latent heat of condensation added to air
Descending air on leeward of mountain
Warm, dry air replaces cooler air

Figure 5.21
Typical conditions which produce a foehn wind.

nized. It is well to keep in mind that the temperature of the foehn or chinook is *relatively* warm, that is, it replaces colder air at the surface. Its actual temperature may occasionally be below freezing, but because it is dry it can remove snow or ice by sublimation; however, if it replaces air colder than itself, it is properly called a chinook.

The danger of lending too much importance to precipitation and to the release of the latent heat of condensation on windward slopes as the cause of the foehn wind can best be illustrated by examining the *stable-air foehn effect.* Consider a mass of cold stable air moving against a mountain range, as might happen in winter when a polar air mass out of Canada travels along the eastern margin of the Rockies. (See Figure 5.22.) It is a characteristic of stable air that in its upper levels the air is potentially warmer, that is, if the upper air is brought down to lower altitudes it will be warmed adiabatically to temperatures greater than the surrounding air in the mass. When stable air lies against a mountain barrier it is this potentially warmer air which spills over. It warms as it descends, and, in the lowlands, has somewhat warmer temperatures than

Figure 5.22
Stable-air foehn induced by mountain barriers.



the surface air in the parent air mass. West of the Rockies it will normally replace warmer air and will therefore be experienced as a cold wave in contrast to the chinook. Nevertheless, the temperatures on the west side of the barrier are not so low as those on the east. The stable-air foehn effect may be reinduced subsequently by the Cascade Range so that that portion of the cold air mass which reaches the Pacific Coast is further warmed. In making regional comparisons of temperature under such conditions it is, of course, necessary to take into account the different altitudes involved and the residual heat of the surface west of the mountains. Both tend to accentuate the temperature differences, although they can hardly be considered a part of the stable-air foehn effect.

In summer the stable-air foehn brings hot, dry air to the Pacific slope when an anticyclone lies inland; this creates an extreme forest fire danger. It functions in the opposite direction when the relatively stable air blowing out of the Pacific high cell moves against the coastal ranges and the Cascades, and the potentially warmer air of the higher levels crosses the barriers first. Upon descending on the lee side, it is warmed adiabatically and thus contributes to the higher temperatures of the interior.

The foehn wind unquestionably occurs in many mountainous areas of the world, but is best known and most easily recognized where high mountain chains lie approximately at right angles to prevailing winds. Similarly, the stable-air foehn effect requires the proper juxtaposition of a mountain barrier and a stable air mass.

A large group of so-called storms are the *dust storms* of dry climates and drought-stricken areas. They are often actually associated with true storm conditions, for example, a mid-latitude cyclone, but all too rarely are they accompanied by subsequent precipitation. Dust storms result from the action of strong winds in picking up loose earth material and carrying it to great heights and perhaps for great distances as well. During the drought years of the early 1930s, dust storms ravaged a large area of the Great Plains of the United States as well as central Australia. Millions of tons of dust are transported by such storms. In desert regions, where there is little protective cover, dust or sand storms are common when sustained high-velocity winds blow.

The much smaller *dust-devil,* or *whirlwind,* is an eddy that forms over hot, dry land as a result of intense local daytime convection and surface friction. It is a fair-weather phenomenon identified by the whirling column of dust, usually not more than a few yards in diameter and a few hundred feet in height. Ordinarily the dust-devil is not destructive, but it moves loose trash and is unpleasant for anyone caught in its path. Its whirling eddy of air operates on a small scale and may turn in either direction in response to the conflicting air currents that set it in motion. Dust-devils probably form in a manner similar to fair-weather waterspouts, but, lacking the added energy due to condensation, they are generally somewhat smaller.

Electrical storm is a term applied variously to the lightning associated with thunderstorms and to disturbances in the outer atmosphere which affect radio transmission. The latter are frequently caused by displays of the aurora borealis or aurora australis, which in turn seem to show increased activity with the appearance of sunspots in great numbers. Their effect upon meteorological conditions in the troposphere is not fully understood, but they are believed to influence temperature, pressure, and winds in the stratosphere.

STORM AND AIR MASS OBSERVATIONS

A great deal of the information of value in forecasting weather and for climatic records comes from general storm observations. Details of temperature, pressure, humidity, wind, cloudiness, and so on are essentially the symptoms of the storms; it is their peculiar combination that constitutes a specific disturbance. It is one thing to measure directly the various weather elements but quite another to undertake objective observation of an entire storm. Knowledge of areal extent, height, speed of movement, and intensity is desirable, however. Of these, intensity is the most difficult to express in quantitative terms. In connection with storm research, indices combining two or more measured quantities, such as wind force and rate of precipitation, have been established.

Radar storm-detection techniques make it possible to track clouds and precipitation at distances of 300 km or more, and even to discern the size of water droplets or ice particles in the air (Figure 5.23). As a result, storms can be followed continuously and their evolution traced over areas where satisfactory ground observations are lacking.

An obvious but expensive—and often dangerous—method of storm and air mass observation is aerial reconnaissance. Flights into the upper air often reveal a general storm or an air mass which is not evident from isolated radiosonde and winds-aloft reports. Furthermore, the flight path can be altered to correspond to the location of a storm, and successive flights can follow the storm—an impossible maneuver for either ground stations or ground-based radiosondes. Instrumental and visual observations along weather reconnaissance routes have been especially valuable in weather research and forecasting. In both the Atlantic and Pacific areas, hurricanes (or typhoons) have been studied by means of airplane flights into the storms. Strong aircraft and careful navigation are necessary in the reconnaissance of a hurricane, but the information gained aids in forecasting the future path of the storm as well as in understand-

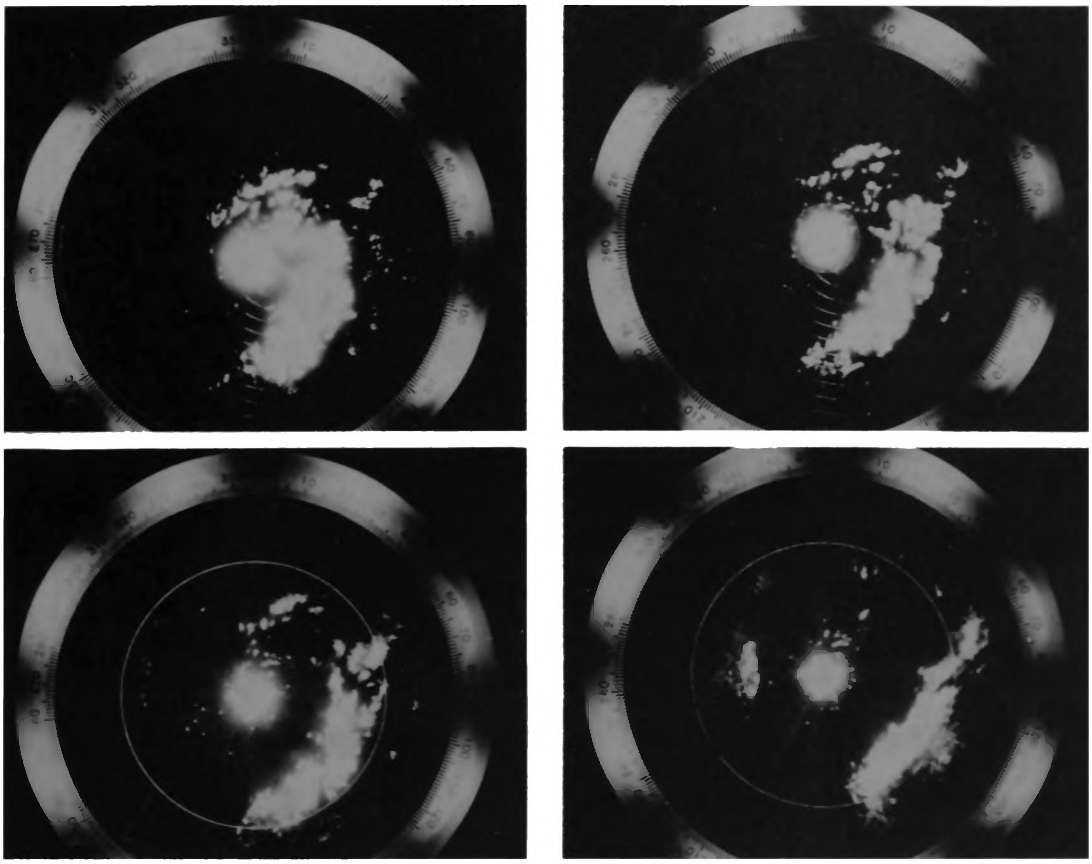


Figure 5.23
Radar images of a moving thunderstorm. Range circles in the lower frames represent a sweep radius of 220 km. Note the new storm forming to the west at 1130 and 1310. (NOAA)

ing the general principles which govern hurricanes. Weather planes are elaborately equipped with meteorological instruments, including radarscopes. Some planes periodically release *dropsondes* to obtain data on conditions in the air below the flight level.

Rockets and satellites, by providing information from ever-greater heights, have opened a new range of techniques in storm observation. Satellites incorporating television and radar that can scan the atmosphere regularly from above furnish data on the extent and movement of storm systems as well as on certain associated phenomena. Polar-orbiting satellites provide photographs of all parts of the earth and its cloud cover (see Figure 5.24). Interpretation of satellite photos by *nephanalysis* (cloud analysis) reveals the development of storms and features of the atmospheric circulation. Weather satellites also facilitate the measurement of radiation, atmospheric temperatures, albedo of the earth and clouds, and the extent of ice and snow cover, all of which have

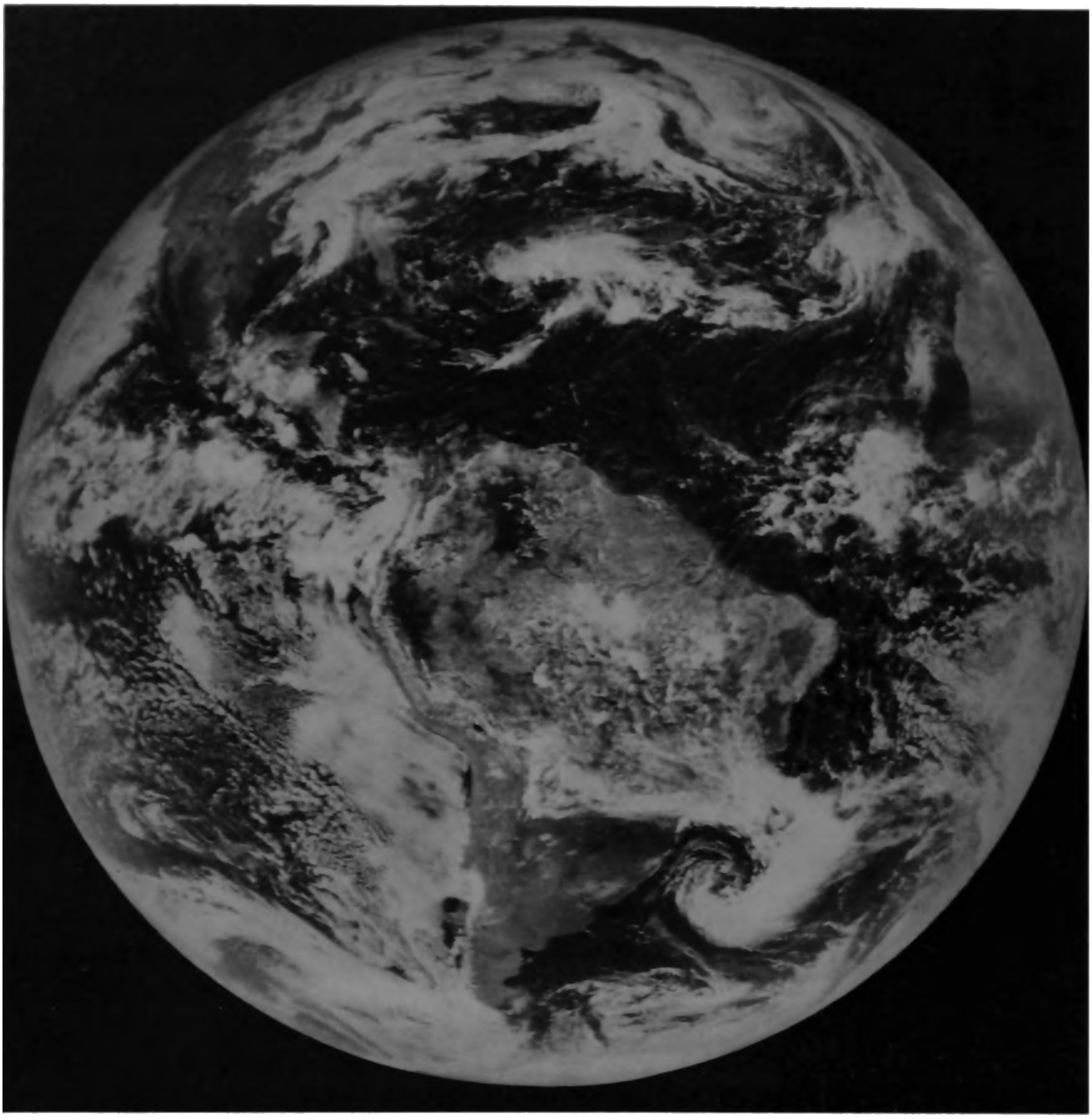


Figure 5.24
Satellite photograph centered on the Amazon Basin. ATS 3 (Applications Technology Satellite) recorded this picture at 1540 G.M.T. on October 13, 1970. Cloud patterns indicate major cyclonic disturbances in the North and South Atlantic. (NOAA, National Environmental Satellite Service)

a direct bearing on weather and climate. Rockets equipped with meteorological sensors extend the range of upper-air observation well beyond that of conventional balloon ascents and thereby provide data at levels intermediate between the latter and satellites.

THE WEATHER MAP

The basic map for display of weather data is the *synoptic weather chart,* which shows the surface distribution of weather elements for a given time. That is, it gives a synopsis of the weather situation. The scale and area covered by the map depend on its intended use. Short-period forecasts for small areas can usually be made from maps covering an area only slightly larger than that included in the forecast. Sectional maps of this type for the local area are commonly used to supplement the larger synoptic map. Forecasts for a large area (such as a continent) or for a longer period require maps that extend well beyond the area in question, preferably to encompass an entire hemisphere.

Synoptic observations are taken simultaneously at many stations throughout the world by agreement through the World Meteorological Organization. Times for these observations are 0000, 0600, 1200, and 1800 Greenwich Mean Time. In addition, many airway stations take hourly observations which aid in making short-period forecasts for flights as well as in more detailed analysis of trends in the weather. Collection of weather data at forecasting stations requires efficient communications, including telephone, telegraph, teletype, radio, and radioteletype. Because a great deal of information must be transmitted in a short time, the observations are sent in the form of a numerical code known as the World Meteorological Organization Code. A sample coded message is shown in Figure 5.25. At the forecasting center the coded messages are decoded and the data are plotted on the synoptic map as figures and symbols around the proper station circles (Figure 5.26). At major forecasting centers computers and automatic plotting devices translate synoptic data directly onto maps, thus reducing the amount of hand plotting that would otherwise be necessary.

On the plotted map the analyst draws in lines to represent fronts, and isobars to indicate pressure distribution (Figure 5.27). Precipitation areas are shaded, and isotherms or other special information may be added. Other aids to forecasting, such as abbreviations of air mass types,

SYMBOLIC FORM OF MESSAGE

iii	Nddff	VVwwW	PPPTT	$N_hC_LhC_MC_H$	T_dT_dapp	$7RRR_ts$

SAMPLE CODED MESSAGE

405	83220	12716	24731	67292	30228	74542

Figure 5.25
Symbolic form of synoptic weather message and sample coded message. (NOAA)

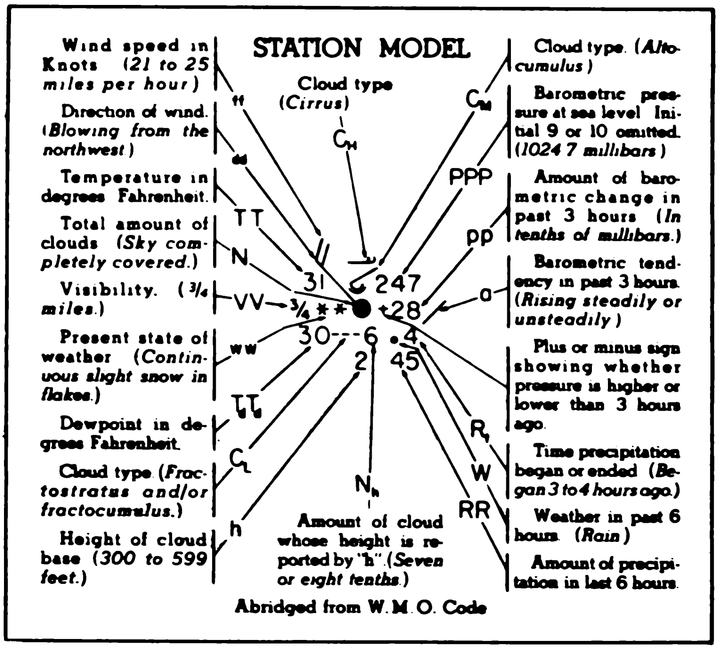

STATION MODEL
Wind speed in Knots (21 to 25 miles per hour)
Direction of wind. (Blowing from the northwest)
Temperature in degrees Fahrenheit.
Total amount of clouds (Sky completely covered.)
Visibility. (3/4 miles.)
Present state of weather (Continuous slight snow in flakes.)
Dewpoint in degrees Fahrenheit.
Cloud type (Fractostratus and/or fractocumulus.)
Height of cloud base (300 to 599 feet.)
ff
dd
TT
N
VV
ww
TdTd
CL
h
Cloud type (Cirrus)
CH
31
3/4
30
6
2
247
+28
4
45
Nh
Amount of cloud whose height is reported by "h". (Seven or eight tenths.)
CM
PPP
pp
a
R1
W
RR
Cloud type. (Altocumulus)
Barometric pressure at sea level. Initial 9 or 10 omitted. (1024.7 millibars)
Amount of barometric change in past 3 hours (In tenths of millibars.)
Barometric tendency in past 3 hours. (Rising steadily or unsteadily)
Plus or minus sign showing whether pressure is higher or lower than 3 hours ago.
Time precipitation began or ended (Began 3 to 4 hours ago.)
Weather in past 6 hours (Rain)
Amount of precipitation in last 6 hours.
Abridged from W.M.O. Code

Figure 5.26
Station model for plotting synoptic data on United States weather maps. (NOAA)

are entered on the map after reference to upper-air charts. Facsimile recorders receive maps and charts that are prepared and transmitted by main analysis centers (Figure 5.28).

REGIONAL WEATHER PATTERNS

The world's weather at any instant is a kaleidoscope of many motion systems, each incorporating a unique combination of atmospheric conditions. As time passes each system changes in response to internal dynamic processes, exchanges of energy and moisture with the earth's surface, and interaction with other systems. Although these modifications often appear to be random in nature, they are phenomena that tend to follow regional patterns with some regularity. Thus, it is common to identify tropical, monsoonal, arctic, and other weather types. Over a period of years the recurrence of characteristic weather sequences creates distinctive climates. The chapters of Part Two will treat regional climates of the world.

Weather patterns are of more immediate significance in forecasting, which is concerned with predicting their changing character. Knowledge of regularities, probabilities, and anomalies in weather constitutes the climatological aspect of forecasting. To the extent that weather sequences repeat themselves analogous patterns in the past can assist prediction.

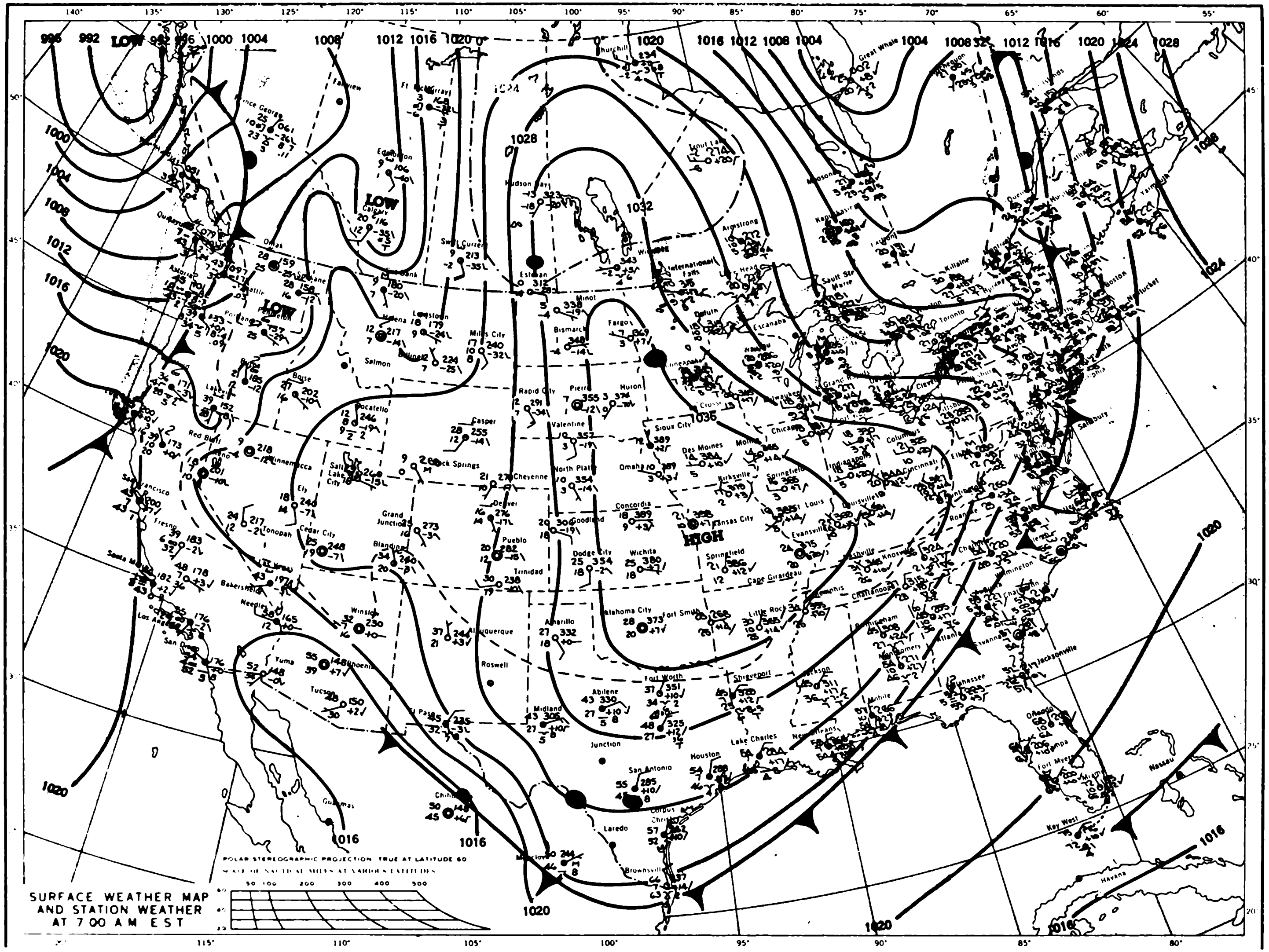


Figure 5.27
Daily Weather Map, November 7, 1971, 7:00 A.M. *E.S.T.* (NOAA, Environmental Data Service)

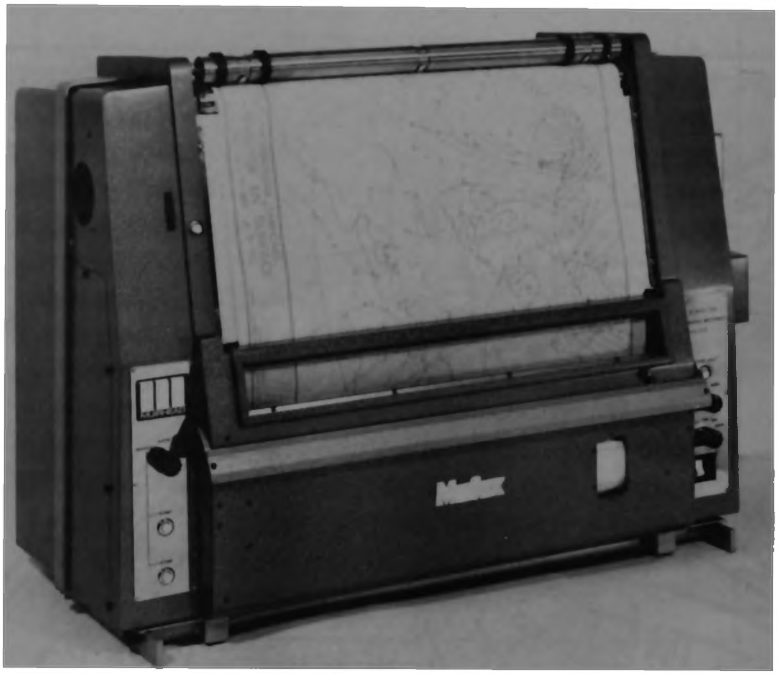


Figure 5.28
Facsimile recorder receiving a surface weather map. (Courtesy Muirhead, Inc.)

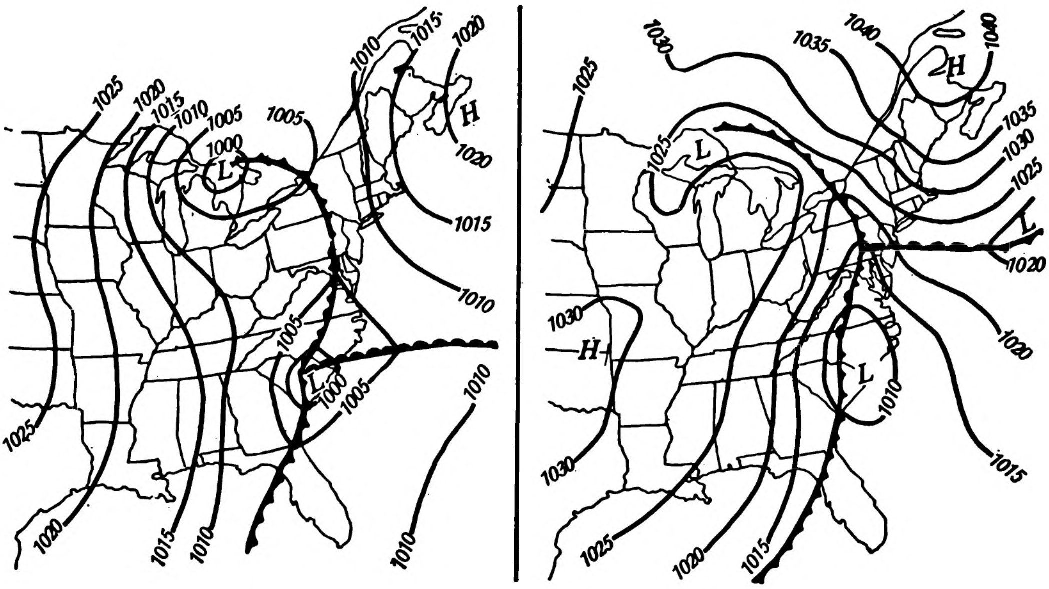


Figure 5.29
Synoptic analogs, Eastern United States. The simplified weather map on the left shows the pressure pattern and fronts on November 8, 1913; the map on the right is for November 25, 1950. An intense storm and heavy snow accompanied both of these analogous cyclonic systems. (After Mook and Smith, from *Water, 1955 Yearbook of Agriculture*)

This principle has enabled countless "old-timers" to achieve remarkably accurate short-term predictions of local weather. The more formal *analog method* of forecasting depends on reference to historical weather maps and records (Figure 5.29). Classification of weather types aids the forecaster's selection of similar weather patterns from the past and examination of the succession of weather as a guide to determining future trends.

Four distinct phases of activity lead to the modern meteorological forecast: observation, communication, analysis, and prognosis. The first two are obviously necessary to provide data that reveal existing conditions, which soon become a record of the past. Analysis entails the mathematical-physical interpretation of observational data and is facilitated by graphic displays on maps and charts as well as by computers. Identification of such phenomena as pressure systems, air masses, or fronts and determination of their relationship to major circulation patterns are examples of the analytical phase of weather forecasting. Ideally, the prognosis should evolve from a complete understanding of atmospheric processes; this would enable the forecaster to predict the movement of motion systems and their associated weather. Unfortunately, neither the structure nor the application of the complex equations that express exchanges of moisture, heat, and momentum have been sufficiently refined to ensure the desired level of accuracy. Numerical forecasting techniques, with the aid of computers, have greatly accelerated the solution of mathematical problems involving simultaneous variables in a three-dimensional atmosphere. Basic research on the physics of the atmosphere will enhance further improvements in objective forecasting.

Additional Readings

BATTAN, LOUIS J., *The Nature of Violent Storms.* New York: Anchor Books, Doubleday & Company, Inc., 1961.

CHANGNON, STANLEY A., JR., and JOHN W. WILSON, "Record Severe Storms in a Dense Meteorological Network," *Weatherwise,* 24, 4 (1971), 152–63.

COOK, A. W., and A. G. TOPIL, "Some Examples of Chinooks East of the Mountains in Colorado," *Bull. Am. Met. Soc.,* 33, 2 (1952), 42–47.

FUJITA, T. THEODORE, "Tornadoes Around the World," *Weatherwise,* 26, 2 (1973), 56–62, 78–83.

KESSLER, EDWIN, "Tornadoes," *Bull. Am. Met. Soc.,* 51, 10 (1970), 926–36.

NEWTON, C. W., "Severe Convective Storms," *Adv. in Geophys.,* 12 (1967), 257–308.

NORDBERG, W., and HARRY PRESS, "The Nimbus I Meteorological Satellite," *Bull. Am. Met. Soc.,* 45, 11 (1964), 684–87.

PEARSON, A. D., and A. F. SADOWSKI, "Hurricane-induced Tornadoes and Their Distribution," *Mon. Wea. Rev.,* 93, 7 (1965), 461–64.

PYLE, ROBERT L., "Weather Satellite Capabilities: Present and Future," *Weatherwise,* 25, 5 (1972), 208–19.

RAMAGE, C. S., "The Subtropical Cyclone," *Journ. Geophys. Res.,* 67, 4 (1962), 1401–11.

RIEHL, HERBERT, "On the Origin and Possible Modification of Hurricanes," *Science,* 141, 3585 (1963), 1001–10.

SIMPSON, R. H., "Hurricane Prediction: Progress and Problem Areas," *Science,* 181, 4103 (1973), 899–907.

STOMMEL, HENRY, "The Circulation of the Abyss," *Scient. Am.,* 199, 1 (1958), 85–90.

TYSON, P. D., "Berg Winds of South Africa," *Weather,* 19, 1 (1964), 7–11.

PART TWO

Climates of the World

chapter six

Climatic Classification

The many variations in climate from place to place, as determined by different combinations of atmospheric processes, produce a correspondingly large number of climatic types. An area of the earth's surface over which the combined effects of climatic factors result in an approximately homogeneous set of climatic conditions, that is, a climatic type, is termed a *climatic region*. To facilitate description and mapping of climatic regions it is necessary to identify and classify the respective types. Regional climatology is concerned with this task and makes use of analytical and descriptive techniques in its search for order in the world climatic pattern.

APPROACHES TO CLIMATIC CLASSIFICATION

As a fundamental tool of science, classification has three interrelated objectives: to bring order to large quantities of information, to speed retrieval of information, and to facilitate communication. Classification of climate shares these objectives. It is concerned with organization of climatic data in such a way that both descriptive and analytical generalizations can be made, and it attempts to store information in an orderly manner for easy reference and communication, often in the form of maps. The value of a systematic arrangement of climates is determined largely by its intended use; a system that suits one purpose is not necessarily useful for another. For example, a classification based on critical temperature and moisture limits for growth of a certain plant or animal organism might serve the needs of a biological study, but it is not likely to be satisfactory for weather forecasting, which relies more on such factors as the general circulation, storm types, and weather probabilities. Thus, in the design of a climatic classification we should begin by defining the purpose. Three broad approaches are equally feasible: (1) empirical, (2) genetic, and (3) applied. Together they constitute a classification of classifications, but the features of all three may be incorporated in a single system.

Empirical classifications are based on the observable features of climate, which may be treated singly or in combination to establish criteria for climatic types. Temperature criteria, for example, might yield "hot," "warm," "cool," and "cold" climates, each of which can be defined in terms of strict mathematical limits. Adding precipitation and other elements to the criteria, the number of possible combinations rapidly multiplies, and soon the system becomes unwieldy. It is, therefore, necessary to select the criteria that are most significant in light of the intended purpose. Heat and moisture factors have dominated empirical classification, but all elements are inherently significant for one purpose or another.

Genetic classification attempts to organize climates according to their causes. Ideally, the criteria employed in the differentiation of climatic types should reflect their origins if climatology is to be explanatory as well as descriptive. In practice, however, explanations are often theoretical, incomplete, and difficult to quantify. Genetic classification also is subject to theoretical biases; a system based on causes tends to perpetuate faulty or over-generalized theories. The ancient Greeks recognized a relationship between latitude and temperature and devised a system of *klimata,* or zones (torrid, temperate, and frigid), that have persisted in writings to the present day in spite of evidence that net radiation does not vary solely with latitude and that other factors affect the distribution of temperature. Besides latitude, features of the general circulation, effects of oceans and continents, mountain barriers, and altitude are other common bases for genetic classification, producing such types as polar and tropical, tradewind littoral, maritime and continental, or highland climates.

Applied (also known as technical or functional) classifications of climate assist in the solution of specialized problems that involve one or more climatic factors. They define class limits in terms of the effects of climate on other phenomena. Outstanding among modern attempts at climatic classification are those that seek a systematic relationship between climatic factors and the world pattern of vegetation. Natural vegetation integrates the effects of climate better than any instrument that has so far been designed, and it is thus an index of climatic conditions. By referring to the major plant associations, biologists have tried to determine the climatic factors that correspond with areal differences in vegetation. Numerous correlations between vegetation and heat or moisture factors have been discovered, permitting the use of temperature or moisture indices as criteria for climatic types. The resulting types and their regional boundaries approach reality in terms of the associated vegetation, while retaining a climatic basis. Commonly, classifications of this kind employ vegetation terms. Rain forest, desert, steppe, and tundra are names that have a climatic connotation. In each case there are climatic limits beyond which the characteristic plant association (or a specific indicator species) does not occur naturally. Fluctuations in climate and in nonclimatic influences lead to problems in the delineation of boundaries, however.

An improvement upon the use of simple rainfall or temperature limits in climatic classification recognizes the relationship between heat and moisture factors. Under high temperatures plants require more precipitation to meet the needs of evapotranspiration. Rains totalling 25 cm annually may support but little plant life in a hot, tropical desert, but that amount may be sufficient for coniferous forest in the cool higher latitudes. Thus natural vegetation is an expression of the adequacy of moisture under a given set of temperature conditions. Precipitation-evaporation ratios and the concept of potential evapotransportation have been introduced to indicate this more complex relationship and establish criteria for climatic types. Variability and seasonal distribution of precipitation and temperature are additional factors which influence plant growth and must be taken into account in any classification that derives from climate-vegetation relationships.

Human health and comfort suggest another possible approach to defining climatic types, with potential applications in clothing design, housing, physiology, and medicine. In much the same way that heat and moisture data are used to determine critical boundaries for natural vegetation or crops, optimum and limiting values of climatic elements afford a basis for classification in terms of human response. Everyone is aware that the reaction of the body to a given air temperature is conditioned by wind, humidity, and sunshine. An individual's state of health, emotional outlook, type of clothing, degree of acclimatization, and a host of other factors also influence personal reaction to climate. One's own reaction to a climate is perhaps a satisfactory basis for a "personal" classification, but as a scientific approach to the problem it is attended by great complexity. The human body appears to be a far less dependable instrument for integrating climatic elements than is a plant. Nevertheless, there have been fruitful studies in the field of applied climatology which have related climate to health, clothing, diet, etc. Maps which show the regional distribution of such relationships are in effect very specialized climatic classifications.

It should be evident that there are many possible classifications of climate, for classification is a product of human ingenuity rather than a natural phenomenon. A complete classification should provide a system of pyramiding categories, ranging from the many microclimates of exceedingly small areas, through mesoclimates, to major macroclimatic divisions on a world scale. But the description of world climates is not easily accomplished as the summation of a great number of microclimates, nor are microclimates easily fitted into the pattern of major world climatic regions. The higher categories of any classification system are necessarily generalizations; the lowest category must include individuals. Great difficulties attend the delimitation of an "individual climate," for climates vary as a continuum over the entire earth. The concepts of *topoclimate,* representing the climate of a spatially finite homogeneous surface, and *ecoclimate,* the climatic environment of a living organism, are useful approximations. In any event, an individual climate is the synthesis of all the climatic elements in a unique combination that results from interacting physical processes. Since the exchanges of energy and mass between the air and the earth's surface are basic climatic processes there is no reason to restrict the study of climate or its classification to the atmosphere.

In order to achieve objectivity in defining the categories of a system it is useful to have quantitative measurements of the climatic elements. In the past the lack of adequate records with respect both to periods covered and to worldwide distribution has presented a serious obstacle. The more than 100,000 surface weather-observing stations of all types in the world are by no means evenly distributed, and many of them record only one or two climatic elements during short or irregular periods. Although 3,500 ships take meteorological observations that are of great value in weather forecasting, the transient nature of these stations limits the use of their records for climatological analyses. The small number of stationary weather ships provide only a token record of climate for huge expanses of water. Eventually these problems should be overcome by satellite observations, which offer the promise of a thorough charting of world climates.

Some knowledge of the common classifications of climate is needed for an understanding of the objectives and methods of regional climatology. The following sections outline the more widely used systems.

KOEPPEN'S CLASSIFICATION

The most used system of classification in its original form or with modifications is that of Wladimir Koeppen (1846–1940), a Russian-born biologist who devoted most of his life to climatic problems. Koeppen aimed at a scheme which would relate climate to vegetation but would provide an objective, numerical basis for defining climate types in terms of climatic elements. Using the world vegetation map of Alphonse de Candolle, a French plant physiologist, he devised his first classification (1900) largely on the basis of vegetation zones and later (1918) revised it with greater attention to temperature, rainfall, and their seasonal characteristics.

The Koeppen system includes five major categories which are designated by capital letters as follows:

A Tropical forest climates; hot all seasons.
B Dry climates.
C Warm temperate rainy climates; mild winters.
D Cold forest climates; severe winters.
E Polar climates.

In order to represent the main climatic types additional symbols are added. Except in the dry climates the second letter refers to rainfall regime, the third to temperature characteristics, and the fourth to special features of the climate. Table 6.1 shows the main climatic types of a modified Koeppen system. Tables 6.2 and 6.3 explain the symbols and boundary criteria used in designating the subdivisions.

In actual application of the system to climatic statistics a great number of subdivisions are possible using the symbols noted in the tables. The numerical limits for certain of the subdivisions are different for each of the higher categories, making the detailed use of the system rather complicated. Several climatologists and geographers have made modifications of the Koeppen classification. The German climatologist, Rudolf Geiger, collaborated with Koeppen on revisions of the world climatic map. He and others continued to modify it after Koeppen's death. It is inevitable that boundary revisions will need to be made as new climatic data become available. One of the best-known modifications of the Koeppen system in the United States is that by Glenn T. Trewartha, who has simplified the world climatic map and redefined several climatic types.

TABLE 6.1*
Main Climatic Types of the Koeppen Classification

Af	Tropical rain forest. Hot; rainy all seasons
Am	Tropical monsoon. Hot; seasonally excessive rainfall
Aw	Tropical savanna. Hot; seasonally dry (usually winter)
BSh	Tropical steppe. Semiarid; hot
BSk	Mid-latitude steppe. Semiarid; cool or cold
BWh	Tropical desert. Arid; hot
BWk	Mid-latitude desert. Arid; cool or cold
Cfa	Humid subtropical. Mild winter; moist all seasons; long hot summer
Cfb	Marine. Mild winter; moist all seasons; warm summer
Cfc	Marine. Mild winter; moist all seasons; short cool summer
Csa	Interior Mediterranean. Mild winter; dry summer; hot summer
Csb	Coastal Mediterranean. Mild winter; dry summer; short warm summer
Cwa	Subtropical monsoon. Mild winter; dry winter; hot summer
Cwb	Tropical upland. Mild winter; dry winter; short warm summer
Dfa	Humid continental. Severe winter; moist all seasons; long, hot summer
Dfb	Humid continental. Severe winter; moist all seasons; short warm summer
Dfc	Subarctic. Severe winter; moist all seasons; short cool summer
Dfd	Subarctic. Extremely cold winter; moist all seasons; short summer
Dwa	Humid continental. Severe winter; dry winter; long hot summer
Dwb	Humid continental. Severe winter; dry winter; warm summer
Dwc	Subarctic. Severe winter; dry winter; short cool summer
Dwd	Subarctic. Extremely cold winter; dry winter; short cool summer
ET	Tundra. Very short summer
EF	Perpetual ice and snow
H	Undifferentiated highland climates

*The data included in Tables 6.1, 6.2, and 6.3 are based on W. Koeppen, *Grundriss der Klimakunde* (Berlin: Walter de Gruyter Company, 1931); W. Koeppen, "Das geographische System der Klimate," Vol I, Part C, of W. Koeppen and R. Geiger, *Handbuch der Klimatologie* (Berlin: Gebruder Borntraeger, 1936); and modifications by R. Geiger, R. J. Russell, Glenn T. Trewartha, and others.

TABLE 6.2

Criteria for Classification of Major Climatic Types in Modified Koeppen System *(Based on mean annual and mean monthly values of precipitation in centimeters and temperature in °C)*

Letter symbol			*Explanation*
1st	*2nd*	*3rd*	
A			Average temperature of coolest month 18°C or higher
	f		Precipitation in driest month at least 6 cm
	m		Precipitation in driest month less than 6 cm but equal to or greater than $10 - r/25$
	w		Precipitation in driest month less than $10 - r/25$
B			70% or more of annual precipitation falls in warmer six months (April through September in the Northern Hemisphere) and r less than $2t + 28$ OR 70% or more of annual precipitation falls in cooler six months (October through March in Northern Hemisphere) and r less than $2t$ OR Neither half of year with more than 70% of annual precipitaiton and r less than $2t + 14$
	W		r less than $\frac{1}{2}$ upper limit of applicable requirement for B
	S		r less than upper limit for B but more than $\frac{1}{2}$ that amount
		h	t greater than 18°C
		k	t less than 18°C
C			Average temperature of warmest month greater than 10°C and of coldest month between 18° and 0°C
	s		Precipitation in driest month of summer half of year less than 4 cm and less than $\frac{1}{3}$ the amount in wettest winter month
	w		Precipitation in driest month of winter half of year less than $\frac{1}{10}$ of amount in wettest summer month
	f		Precipitation not meeting conditions of either s or w
		a	Average temperature of warmest month 22°C or above
		b	Average temperature of each of four warmest months 10°C or above; temperature of warmest month below 22°C
		c	Average temperature of from one to three months 10°C or above; temperature of warmest month below 22°C
D			Average temperature of warmest month greater than 10° and of coldest month 0°C or below
	s		Same as under **C**
	w		Same as under **C**
	f		Same as under **C**
		a	Same as under **C**
		b	Same as under **C**
		c	Same as under **C**
		d	Average temperature of coldest month below −38°C (**d** is then used instead of **a**, **b**, or **c**)
E			Average temperature of warmest month below 10°C
	T		Average temperature of warmest month between 10° and 0°C
	F		Average temperature of warmest month 0° or below
H			Temperature requirements same as **E**, but due to altitude (generally above 1,500 m)

*In formulas t is the average annual temperature in °C; r is average annual precipitation in centimeters.

TABLE 6.3
Supplementary Subdivision Symbols and Criteria in Koeppen Classification

g	Ganges type of temperature regime; maximum before the summer rainy season
i	Annual temperature range less than 5°C
k'	Same as k but average temperature of warmest month less than 18°C
l	Mild; average temperature in all months between 10° and 22°C
n	Frequent fog
n'	Infrequent fog; high humidity, low rainfall, warmest month temperature below 23°C
p	Same as n', except warmest month between 23° and 28°C
p'	Same as n', except warmest month above 28°C
u	Coolest month after summer solstice
v	Warmest month in autumn
w'	Rainy season in autumn
w''	Two distinct rainfall maximums separated by two dry seasons
x	Maximum rainfall in spring or early summer, dry in late summer
x'	Same as x but with infrequent heavy rains in all seasons

THE THORNTHWAITE CLASSIFICATIONS

The American climatologist C. Warren Thornthwaite (1899–1963) introduced his first classification of climates in 1931, when he published a climatic map of North America; in 1933 he extended the system to the world. Its essential features are a *precipitation effectiveness* index, which relates precipitation to values of evaporation derived from temperature data, and a *temperature efficiency* index based on mean monthly temperatures. Subdivisions provide for seasonal distribution of the basic index values, yielding 32 world climatic types. The system maintains a link with that of Koeppen: it expresses the relationship between climate and vegetation in five major humidity provinces: rain forest, forest, grassland, steppe, and desert.

In 1948 Thornthwaite proposed as a basis for climatic classification the concept of *potential evapotranspiration,* that is, the amount of moisture that would be evaporated from the soil and transpired from vegetation if it were available. He regarded potential evapotranspiration as a climatic factor equal in importance to precipitation. If more water were available in a hot desert there would be more vegetation, and more water would be used in evaporation and transpiration. The water need generated by available energy is greater in summer than in winter and greater in hot climates than cold. By comparing the amount of water available from precipitation with the water need it is possible to assess the moisture conditions to determine seasonal surpluses or deficiencies and whether a climate is truly wet or dry. (See Figure 6.1.) Inasmuch as potential evapotranspiration represents a transfer of both heat and

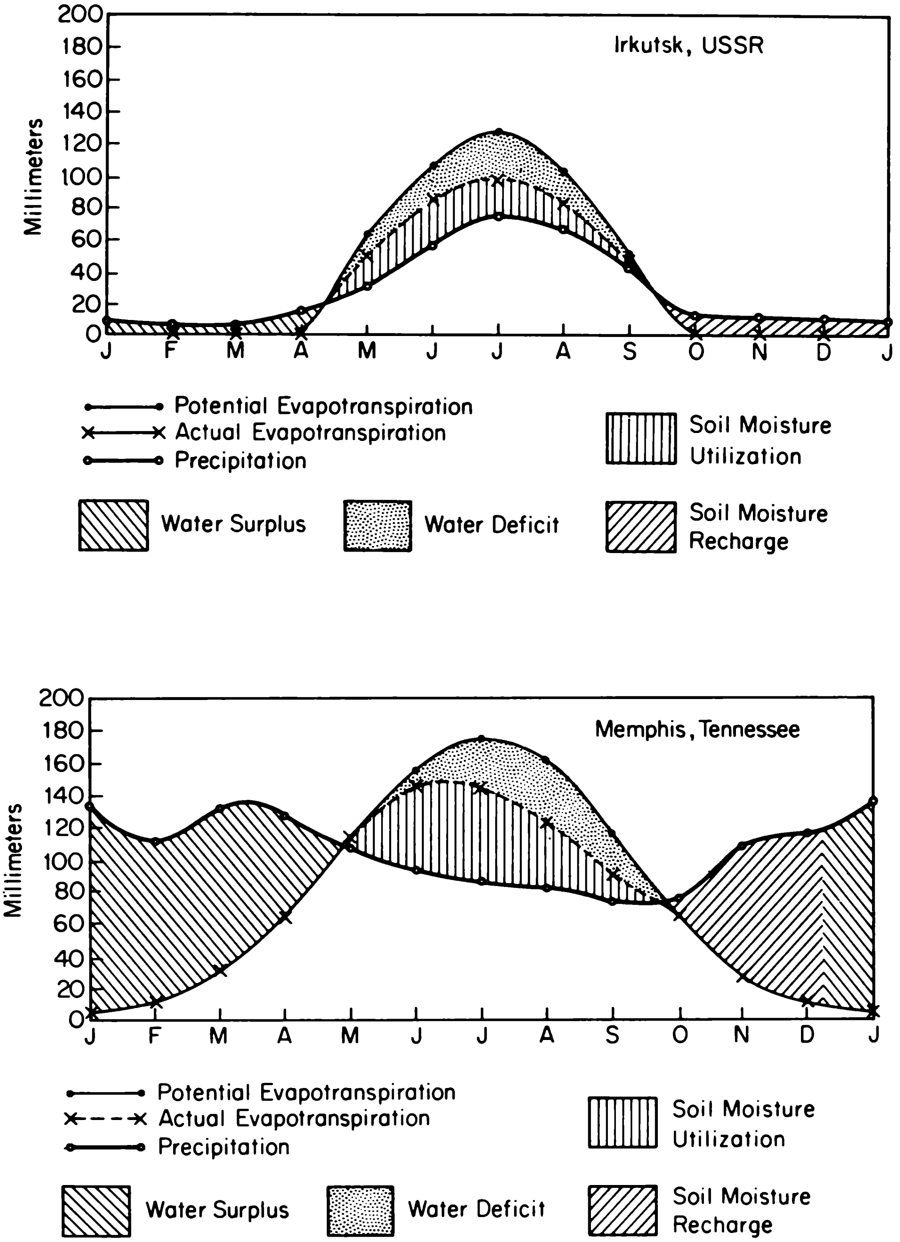


Figure 6.1

Water budget graphs for Irkutsk and Memphis. (After method of Thornthwaite, based on data computed by C. W. Thornthwaite Associates, Laboratory of Climatology)

moisture to the atmosphere and is primarily a function of energy received from the sun, it is an index of *thermal efficiency* as well as water loss, and so combines both the moisture and heat factors in climate. Thus, a climate may be considered as the existing balance between incoming and outgoing heat and moisture at the earth's surface.

Actual measurements of potential evapotranspiration are inadequate in number and duration for a worldwide classification of climates. For calculating values Thornthwaite devised a complex empirical formula in which temperature and daylight hours (a function of latitude) are the variables. (Other methods of determining potential evapotranspiration will be treated in Chapter 11.) In the classification as slightly revised in 1955 potential evapotranspiration (*PE*) and precipitation (*P*) are the bases of four climatic criteria: *moisture adequacy, thermal efficiency, seasonal distribution of moisture adequacy,* and *summer concentration of thermal efficiency.* Each is represented by an index value, and boundaries are set quantitatively. When adjustments are made for the storage of water in the soil the difference between mean monthly amounts of *P* and *PE* is the monthly surplus (*S*) or deficit (*D*). Moisture adequacy may then be expressed by the monthly *moisture index* (I_m) in the formula:

$$I_m = 100 \frac{(S - D)}{PE}$$

If soil moisture is assumed to be constant the equation is simply:

$$I_m = 100 \frac{P}{PE} - 1$$

The sum of the 12 monthly values of I_m gives the annual moisture index. Table 6.4 shows the annual index limits for nine climatic moisture types.

Thermal efficiency is simply the potential evapotranspiration in centimeters. Again, the annual index is the total of monthly values. Note

TABLE 6.4
Climatic Moisture Types*

	Type	*Moisture Index*		
A	Perhumid	100 and above		
B_4	Humid	80	to	100
B_3	Humid	60	to	80
B_2	Humid	40	to	60
B_1	Humid	20	to	40
C_2	Moist subhumid	0	to	20
C_1	Dry subhumid	−33.3	to	0
D	Semiarid	−66.7	to	−33.3
E	Arid	−100	to	−66.7

*Data in Tables 6.4, 6.5, and 6.6 are adapted from Douglas B. Carter and John R. Mather, "Climatic Classification for Environmental Biology," *Publications in Climatology,* 29, 4 (1966).

TABLE 6.5
Thermal Efficiency and its Summer Concentration

Thermal Efficiency			*Summer Concentration*	
TYPE		INDEX (cm)	TYPE	CONCENTRATION (%)
A'	Megathermal	114 and above	a'	below 48.0
B'_4	Mesothermal	99.7 to 114.0	b'	48.0 to 51.9
B'_3	Mesothermal	85.5 to 99.7	b'_3	51.9 to 56.3
B'_2	Mesothermal	71.2 to 85.5	b'_2	56.3 to 61.6
B'_1	Mesothermal	57.0 to 71.2	b'_1	61.6 to 68.0
C'_2	Microthermal	42.7 to 57.0	c'_2	68.0 to 76.3
C'_1	Microthermal	28.5 to 42.7	c'_1	76.3 to 88.0
D'	Tundra	14.2 to 28.5	d'	above 88.0
E'	Frost	below 14.2		

that it is an index of energy in terms of water depth, but that it is derived from thermal data. Table 6.5 gives the nine thermal efficiency types.

Indices of aridity and humidity are used to determine the seasonal distribution of moisture adequacy. In moist climates annual water deficit taken as a percentage of annual *PE* gives the *aridity index.* In dry climates water surplus as a percentage of annual *PE* gives the *humidity index.* Inspection of monthly data reveals the season of surplus or deficit (see Table 6.6).

The percentage of the mean annual *PE* that accumulates in the three summer months is the summer concentration of thermal efficiency (see Table 6.5). By combining the four elements of the classification it is possible to represent the climate of a place by four letter symbols (see examples in Table 6.7). The complexity of the system has made it difficult to display the great number of different climates cartographically, al-

TABLE 6.6
Seasonal Moisture Adequacy

Moist Climates (A, B, C_2)		*Aridity Index*
r	little or no water deficit	0 to 10
s	moderate summer deficit	10 to 20
w	moderate winter deficit	10 to 20
s_2	large summer deficit	above 20
w_2	large winter deficit	above 20
Dry Climates (C_1, D, E)		*Humidity Index*
d	little or no water surplus	0 to 16.7
s	moderate winter surplus	16.7 to 33.3
w	moderate summer surplus	16.7 to 33.3
s_2	large winter surplus	above 33.3
w_2	large summer surplus	above 33.3

TABLE 6.7
Thornthwaite Classification of Selected Stations

Annual Water Budget Criteria	*Alice Springs, Australia ($EA'da'$)*	*San Francisco, California ($C_1 B'_1 da'$)*	*Moscow U.S.S.R. ($C_2 C'_2 rb'_1$)*
Precipitation (cm)	24.5	55.1	63.1
Thermal efficiency (*PE* in cm)	116.2	70.2	55.4
Summer concentration (% of *PE*)	43.8	33.3	64.4
Surplus (cm)	0	3.6	10.4
Deficit (cm)	91.7	18.7	4.5
Humidity index (%)	0	5.1	18.8
Aridity index (%)	78.9	26.6	8.1
Moisture index	−78.9	−21.5	10.7

though continental maps of the separate elements have been prepared. The most significant applications of the potential evapotranspiration concept have been in practical studies of the water budget in relation to water use. Chapters 11 and 12 will treat examples.

OTHER CLASSIFICATIONS

Among the many classifications of world climates most have been general and qualitative, especially when displayed on maps; usually they rely heavily on one or two climatic elements or genetic factors. They make little provision for distinguishing climates in the lower categories or for mathematical determination of climatic boundaries. Geographers have developed several useful world classifications with the primary objective of establishing world regions as a framework for study of human settlement and land-use patterns. Applied classifications based on human responses to climate have been refined progressively in recent years.

Some authors have made valuable contributions to the description of world climates without systematic classification of climatic types. Notable among them was W. G. Kendrew, a British climatologist, who described climates of the continents in terms of the distribution of elements and related genetic factors. His treatment delineates no specific climatic boundaries or types, but it subdivides the continents into geographic regions for the purposes of organized description.

CLIMATIC REGIONS OF THE WORLD

Inasmuch as the world distribution of climatic types is primarily the result of heat and moisture regimes, we can classify climate in broad categories based on the interchange of heat and moisture between the surface and overlying air masses, which tend to dominate the climates of different regions. The climatic types thus defined are: (I) Climates dominated by equatorial and tropical air masses; (II) climates dominated by tropical and polar air masses; and (III) climates dominated by polar and arctic-type air masses. The final group in this basic category is (IV) the highland climates, which have distinctive features arising from the effects of altitude. Subdivisions of these four climatic groups into climatic types is based upon the regional distribution of climatic elements—especially temperature and precipitation—and their seasonal variations. The following outline lists the principal climatic types to be considered in the remaining chapters of Part Two. It follows the customary organization of world climates, beginning at the equator and proceeding toward the poles.

I. *Climates dominated by equatorial and tropical air masses*
   1. Rainy tropics
   2. Monsoon tropics
   3. Wet-and-dry tropics
   4. Tropical arid climate
   5. Tropical semiarid climate

II. *Climates dominated by tropical and polar air masses*
   6. Dry summer subtropics
   7. Humid subtropics
   8. Marine climate
   9. Mid-latitude arid climate
   10. Mid-latitude semiarid climate
   11. Humid continental warm summer climate
   12. Humid continental cool summer climate

III. *Climates dominated by polar and arctic-type air masses*
   13. Taiga
   14. Tundra
   15. Polar climate

IV. *Climates having altitude as the dominant control*
   16. Highland climates

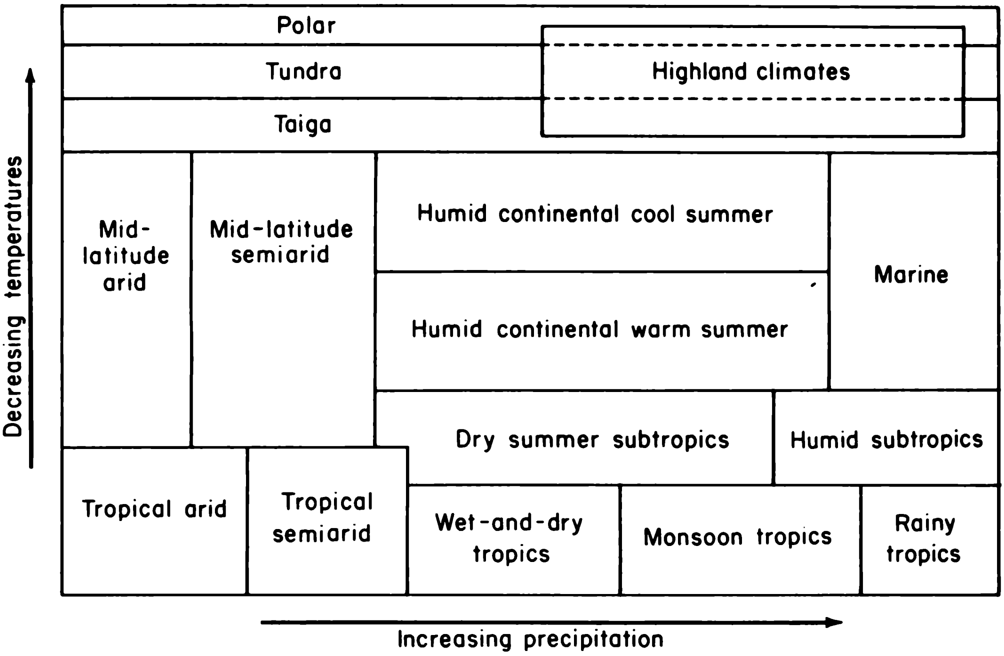


Figure 6.2
Schematic relationship of major climatic types to temperature and precipitation.

This system of climatic types is designed to facilitate the explanatory description of world climates. It will be evident that it follows the Koeppen classification in its basic approach to major categories, relying heavily on temperature and precipitation, their seasonal distribution, and their relation to natural vegetation as criteria for the climatic types. The schematic relationship of climatic types to the temperature and moisture factors is diagrammed in Figure 6.2.

This system is in no sense a rational classification arrived at by mathematical computation. Nor is it readily susceptible to refinement to define the many climatic subtypes. Rather, it is intended to aid our examination of the major climates of the earth while introducing a minimum of departure from the terminology one is likely to encounter in readings in geography, climatology, or other sciences dealing with the world patterns of spatial distribution. Classifications of this kind have been widely used by geographers and climatologists, and though they may not always be in agreement as to specific boundaries and subdivisions, the fundamental pattern of climates recurs in all for the reason that a fairly well-ordered system of climates does exist on the earth.

Additional Readings

BAILEY, HARRY P., "Toward a Unified Concept of the Temperate Climate," *Geog. Rev.*, 54, 4 (1964), 516–45.

BROOKS, C. E. P., "Classification of Climates," *Met. Mag.*, 77, 911 (1948), 97–101.

CARTER, DOUGLAS B., and JOHN R. MATHER, *Climatic Classification for Environmental Biology*, Publications in Climatology, XIX, 4. Elmer, N.J.: Laboratory of Climatology, 1966.

CASTELLI, JOSEPH, "A Flow Chart for Climate Classification," *Journ. of Geog.*, 63, 1 (1964), 19–23.

CHANG, JEN-HU, "An Evaluation of the 1948 Thornthwaite Classification," *Ann. Assoc. Am. Geogrs.*, 49, 1 (1959), 24–30.

HARE, F. KENNETH, "Climatic Classification," Chapter VII in *London Essays in Geography*, eds. L. D. Stamp and S. W. Wooldridge. Cambridge, Mass.: Harvard University Press, 1951, pp. 111–34.

MAUNDER, W. J., "A Human Classification of Climate," *Weather*, 12, 1 (1962), 3–12.

OLIVER, JOHN E., "A Genetic Approach to Climatic Classification," *Ann. Assoc. Am. Geogrs.*, 60, 4 (1970), 615–37.

TERJUNG, WERNER H., "Physiologic Climates of the Conterminus United States: A Bioclimatic Classification Based on Man," *Ann. Assoc. Am. Geogrs.*, 56, 1 (1966), 141–79.

———, "Toward a Climatic Classification Based on Net Radiation," *Proc. Assoc. Am. Geogrs.*, 2 (1970), 140–44.

THORNTHWAITE, C. W., "The Climates of North America According to a New Classification," *Geog. Rev.*, 21, 4 (1931), 633–55.

———, "The Climates of the Earth," *Geog. Rev.*, 23, 3 (1933), 433–40.

———, "Problems in the Classification of Climates," *Geog. Rev.*, 33, 2 (1943), 233–55.

———, "An Approach Toward a Rational Classification of Climate," *Geog. Rev.*, 38, 1 (1948), 55–94.

TROLL, C., "Climatic Seasons and Climatic Classification," *Oriental Geogr.*, 2, 2 (1958), 141–65.

chapter seven

Climates Dominated by Equatorial and Tropical Air Masses

Climates under the influence of equatorial and tropical air masses experience high temperatures throughout the year. They lie in low latitudes in the zones of the intertropical convergence, the tradewinds, and the subtropic highs, where temperature contrasts between air masses are not great. The year-round net radiation surplus at low latitudes accounts for the major features of this group of climates, yet there are marked variations in seasonal water budgets.

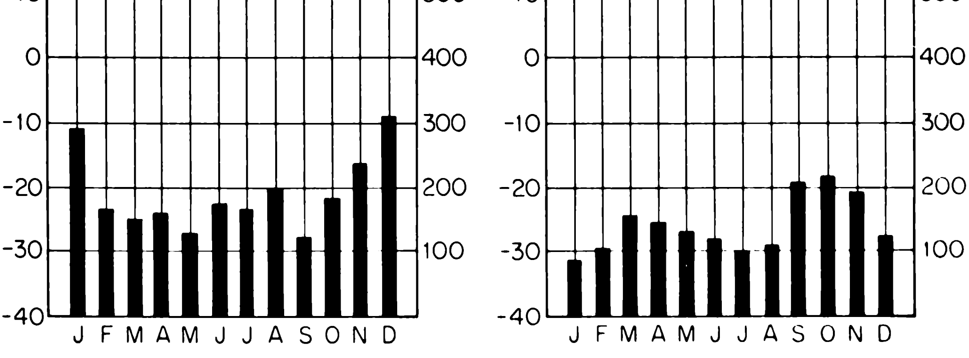
THE RAINY TROPICS

The rainy tropical type of climate prevails in the lowlands on and near the equator and along tropical coasts that are exposed to trade winds but backed by interior highlands. The principal areas of the world having this type of climate are the Amazon Basin of South America, the windward coasts of Central America, the central Congo Basin of Africa, Indonesia, New Guinea, the Philippines, and the east coast of Madagascar. The combination of constantly high temperatures with abundant rainfall well distributed throughout the year makes this a climate literally without seasons. Monthly temperatures average 25° to 28° C, and because there is a net gain of radiation throughout the year the annual range of temperature is small.

In the rainy tropics the diurnal range of temperature is normally much greater than the annual range between the maximum and minimum monthly means. Annual ranges are commonly less than 2 or 3 C°, whereas diurnal ranges may be 8 or 10 C°. Nighttime cooling of moist air under clear skies produces saturation and copious dew or perhaps fog, for the relative humidity is always high. These condensation products evaporate into the warmed air shortly after sunup. Average daytime maximums are usually below 32°C. Singapore has mean daily maximums ranging between 30 and 32° C. Extreme maximums in the rainy tropics are lower than those of many mid-latitude stations, rarely reaching above 38° C. Another feature of temperatures is the small interdiurnal variation; that is, the change in temperature from day to day is small, making the climate monotonous. Whereas in the higher latitudes the climate has a more or less regular seasonal rhythm, in the rainy tropics the cycle is the day.

Annual precipitation in the rainy tropics exceeds 150 cm at most stations. No month is exceptionally dry, yet a graph of the precipitation regime is by no means as smooth as the annual march of temperature. (See Figure 7.1.) Some stations have a definite maximum of precipitation in one month; a few have regimes with two maximums during the year,

Figure 7.1
Climatic graphs for Singapore and Coquilhatville, rainy tropics.



Singapore	1°N; 18m
Mean annual temperature	27°C
Annual temperature range	2°
Mean annual precipitation	2282 mm

Coquilhatville, Zaire	3°N; 345m
Mean annual temperature	24°C
Annual temperature range	1°
Mean annual precipitation	1678 mm

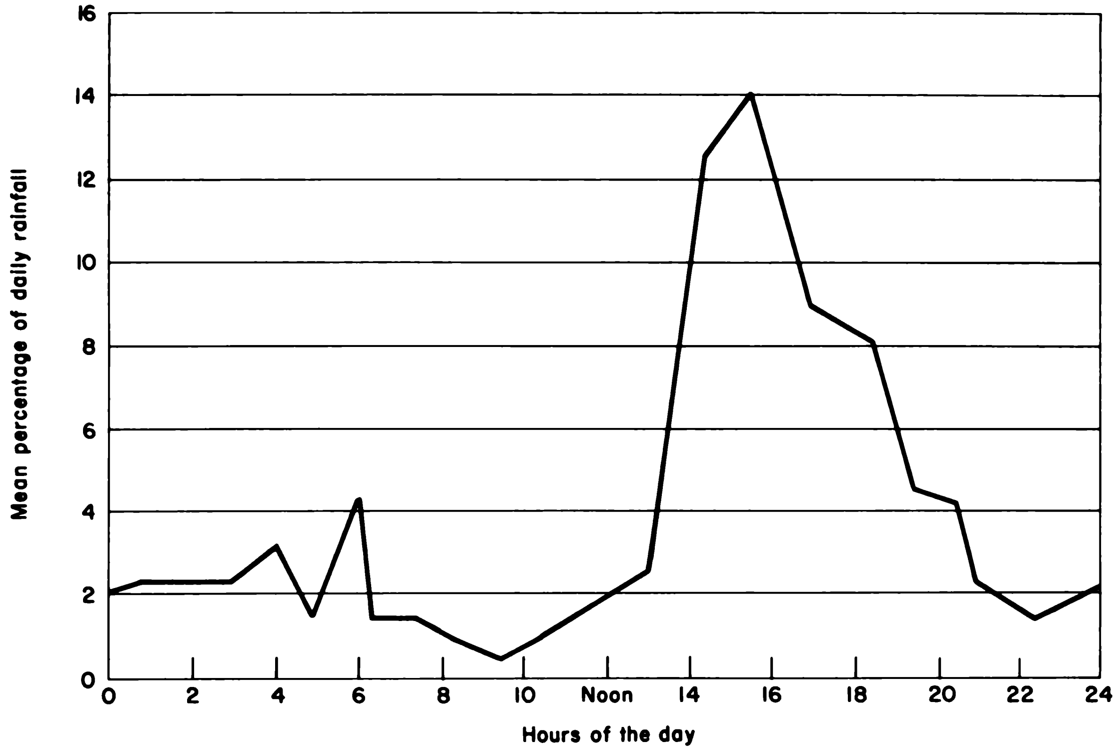


Figure 7.2
Diurnal variation of rainfall at Kuala Lumpur, Malaya. (After Ooi Jin-bee, "Rural Development in Tropical Areas," *Journ. Trop. Geog.,* Vol. 12, March, 1959)

resulting from the seasonal migration of the intertropical convergence. Most rainfall is of the convectional type and comes with thunderstorms. The convergence of moist tropical air and the intense daytime radiation create ideal conditions for thunderstorm formation. Usually the thundershowers are concentrated in small areas and are of short duration. In many areas their maximum occurrence is in the afternoon, and they are frequently preceded and followed by clear, sunny weather. (See Figure 7.2.) Coastal locations and islands near the poleward margins of the rainy tropics may experience hurricanes, but these violent tropical cyclones do not occur along the equator nor in interior areas (Figure 7.3). Hurricanes may account for a considerable proportion of the autumn rainfall of coastal areas visited by this type of storm.

Although thunderstorms are the dominant storm type, shallow cyclonic circulation sometimes develops in the rainy tropics. Two air masses of equatorial or tropical origin are not likely to differ greatly in temperature characteristics, and zones of convergence accordingly tend to be broad troughs of low pressure accompanied by overcast skies and protracted periods of rain. On rare occasions, air masses from middle latitudes invade the tropics and produce frontal rainfall and lowered temperatures, but, for the most part, the storm lines of the tropics which seem to resemble extratropical fronts are zones of convergence within tropical air masses. In this connection it should be remembered that convergence can result from either different wind directions or wind speeds or both. Therefore, a close study of winds is necessary for prediction of weather in the tropics, for tropical weather is primarily of the air mass variety rather than the frontal.

Where mountain ranges lie athwart trade winds which have travelled across warm water, the average annual precipitation of the rainy tropics is greatly exceeded. Examples of these trade-wind coasts are found in the Caribbean Sea, southeastern Brazil, eastern Madagascar, and on the numerous volcanic islands of the tropical Pacific. Wherever the orographic effect is a factor inducing heavy rainfall, leeward sides of the mountains show small annual totals in striking contrast to the heavy rainfall on the windward sides.

Quite as important as the total annual precipitation and its regime is the effectiveness of precipitation. Owing to the constantly high temperatures in the equatorial regions the rates of evaporation and transpiration are high and a correspondingly greater amount of rain is required to maintain satisfactory conditions for plant growth. Thus, the monthly precipitation amounts which might seem adequate in the latitude of the Great Lakes may actually result in drought in equatorial latitudes. Although there is a marked variability of precipitation from year to year in this climate, there are ordinarily no droughts of serious proportions.

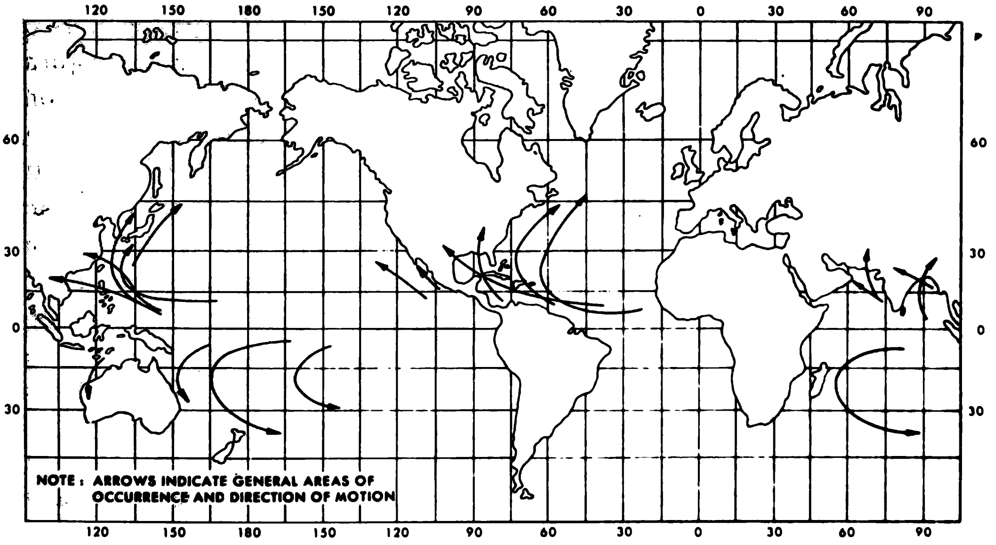

120
150
180
150
120
90
60
30
0
30
60
90
60
30
0
30
NOTE: ARROWS INDICATE GENERAL AREAS OF OCCURRENCE AND DIRECTION OF MOTION

Figure 7.3

Principal world regions of violent tropical cyclones. (Department of the Army)

The air masses which dominate the rainy tropical climate are associated with the subtropic high and with stagnated air in the equatorial convergence. (See Figure 5.4.) The trade winds originate in the subtropical cells of high pressure and flow equatorward. They constitute the most dependable wind systems of this type of climate but are found chiefly along its margins on east coasts. Owing to their effect the rainy tropics reach their farthest poleward extensions. Some authors refer to these areas as the "tropical eastern littorals" or the "tropical windward coasts." Not all trade-wind coasts have the true characteristics of the rainy tropics, however. An example station in the poleward extension of the rainy tropics is Salvador, Brazil, for which climatic data are given in the accompanying table. Along the east coast of Brazil, south of Cape São Roque, the onshore flow of the southeast trades is stronger and more nearly perpendicular to the coast in late autumn and early winter. Salvador thus has a distinct autumn maximum of rainfall rather than the typical dry winter. Suva, Fiji, also has an autumn maximum, partly because of its location in a region visited by the South Pacific subtropic high in winter and spring.

Salvador, Brazil 13°S; 9m

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	26	26	26	26	25	24	23	23	24	24	25	26	25
P (mm)	74	78	163	290	298	195	206	112	85	94	143	98	1837

Suva, Fiji 18°S; 9m

T (°C)	26	26	26	26	25	24	23	23	24	24	25	26	25
P (mm)	321	313	399	385	272	160	162	155	218	216	268	291	3160

The intertropical convergence dominates by far the greater part of the rainy tropics, and associated weather accounts for most tropical precipitation. Because it is essentially a belt of shallow depressions, convection, and convergence, it does not have strong prevailing winds but rather light, erratic air movement except along some trade-wind coasts. The best-developed horizontal temperature gradients in the rainy tropics are between land and sea, where the resulting pressure gradients produce daily land and sea breezes that alleviate daytime heat for many coastal locations.

MONSOON TROPICS

The monsoon tropics are found in close association with the rainy tropics, generally along coasts where there is a seasonal onshore flow of moist air. The principal areas are the west coasts of India and Burma, the east coast of Vietnam, the northern Philippines, the western Guinea Coast of Africa, the northeastern coast of South America, and the northern coasts of Haiti and Puerto Rico. The monsoon tropical climate differs from the rainy tropics in that it has a distinct dry season. However, the storage of soil moisture is sufficient to maintain a forest through the rainless period. It may be regarded as transitional in nature from the rainy tropics to the wet-and-dry tropics, having annual rainfall totals comparable to the former and a regime of precipitation similar to the latter. It is well to remember that the term monsoon tropics does not apply to all climates affected by a monsoonal wind circulation. Its use stems from the characteristic climates of monsoon Asia, but the designation wet-and-dry tropics is applied to those regions with less annual precipitation and a distinct season of drought.

Mean monthly temperatures in the monsoon tropics are not greatly different from those in the rainy tropics—typical values ranging well above 20°C. These climates extend farther poleward, however, and the annual range is greater in some cases. The outstanding feature of the annual march of temperature for a number of stations in the monsoon tropics is the occurrence of the maximum before the high-sun period, that is, in May or June rather than in July in the Northern Hemisphere. Maximum temperatures are usually reached in the period of clearer skies, increasing insolation, and deficient precipitation just before the onset of more persistent cloudiness and heavy rainfall. Trivandrum, India, is a representative station.

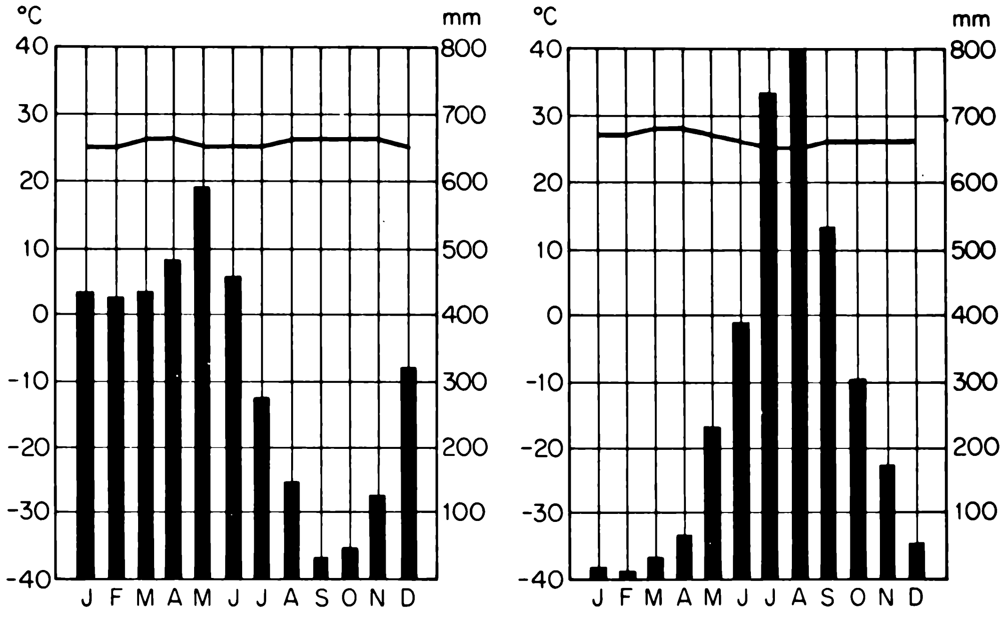
Trivandrum, India 8°N; 64 m

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	27	27	28	28	28	26	26	26	27	27	27	27	25
P (mm)	19	21	44	122	249	331	211	164	123	271	207	73	1,835

Diurnal variation of temperature in the monsoon tropics is, on the average, slightly greater than in the rainy tropics. It is greatest in the drier months and least in the rainy season. During winter there may be some influence from cyclonic disturbances which pass on the poleward margin with resulting short periods of comparatively low temperatures.

Annual precipitation averages above 150 cm in most areas of the monsoon tropics. Where a strong onshore flow of moist air meets a coast backed by a mountain barrier exceptional rainfall totals are achieved. (See Figure 7.4.)

Precipitation which accounts for most of the total in the monsoon tropics is of the showery type, with the orographic effect often playing an important part, but certain areas are visited by tropical depressions. Summer weather, that is, the rainy season, is similar to that of the rainy tropics, usually with greater monthly rainfall totals. Winter is slightly



Cayenne, French Guiana	5°N ; 9m
Mean annual temperature	26°C
Annual temperature range	1°
Mean annual precipitation	3744mm

Freetown, Sierra Leone	9°N; 25m
Mean annual temperature	26°C
Annual temperature range	3°
Mean annual precipitation	3321 mm

Figure 7.4
Climatic graphs for Cayenne and Freetown, monsoon tropics.

cooler and somewhat drier, having protracted sunny periods with occasional thundershowers. Remnants of polar fronts sometimes invade these climates above the level of the trades in winter, producing overcast skies and light rainfall but no very great temperature decreases at the surface.

In southeastern Asia the characteristic pattern of circulation is monsoonal. Air trajectories from the Indian Ocean meet the easterly trades to produce a belt of convergence in summer over the east coast and offshore islands (Figure 7.5). Much of the moisture precipitated along the east Asian littoral is transported aloft from the Indian Ocean. In winter a southern arm of the westerly jet stream lies south of the Himalayan divide, allowing little moisture into the area. Surface winds are from a northerly direction, and upper-level subsidence suppresses vertical cloud development.

In other regions where the monsoon tropical climate prevails, the seasonal variation in strength and persistence of the trades combines with the effect of contrasting land and water surfaces to produce a "monsoon tendency." Although essentially easterly winds, the trades at times play a part in onshore flow of west winds, notably on the Guinea Coast of Africa and the Malabar Coast of India. As the world pattern of the gen-

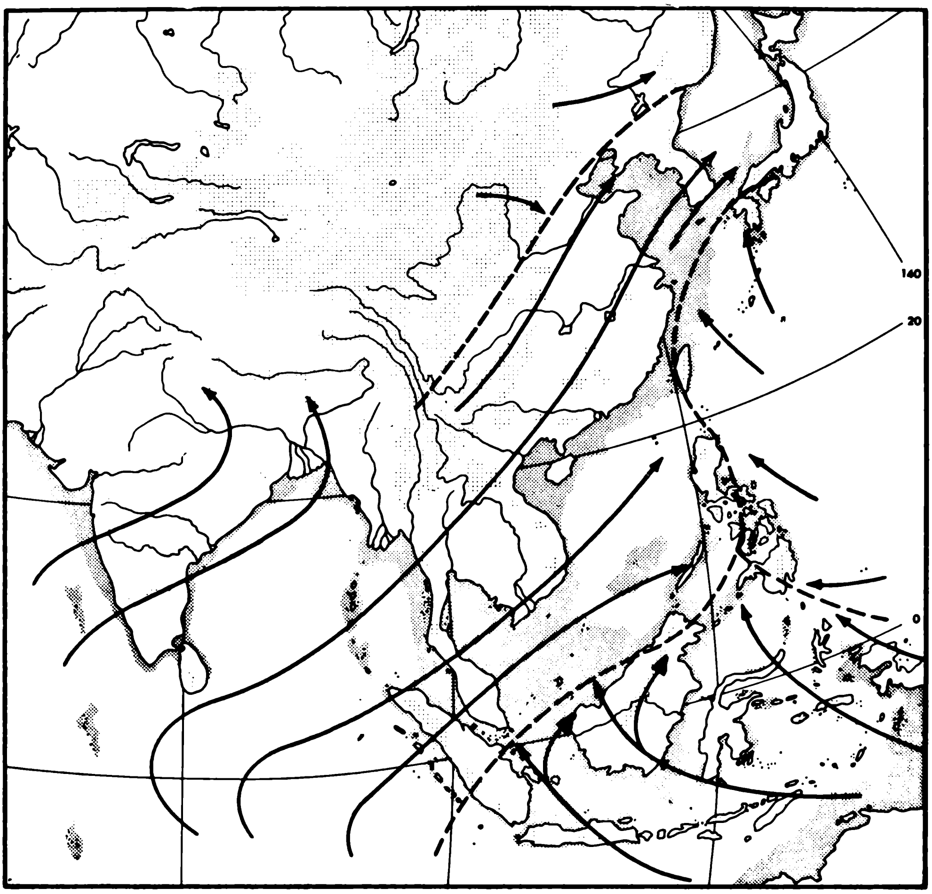


Figure 7.5
Mean air trajectories over Southeast Asia in July. Dashed lines represent zones of convergence. (After Watts, Thompson, and others)

eral circulation migrates northward in the Northern Hemisphere summer the southeast trades take a position astride or even slightly north of the equator. Blowing into the Northern Hemisphere they come under the opposite influence of the Coriolis effect and curve to the right to become southwest winds as they approach these coasts. In the case of India they merge into the monsoon circulation. (See Figure 7.5.)

Wherever the trades are found they vary not only with the seasons, migrating to some extent with the general circulation, but also their strength varies periodically in what has been called the "surge of the trades." Periods of greater velocity in the trades are caused by disturbances from higher latitudes which create a steeper pressure gradient away from the subtropic high centers. When these surges are developing a wind discontinuity, or *shear line,* exists between the strong winds of the surge and the weaker winds of the trades nearer the equator. Convergence along the shear line results in clouds of the cumulus type and showers.

In areas of the monsoon tropics where the monsoon circulation is

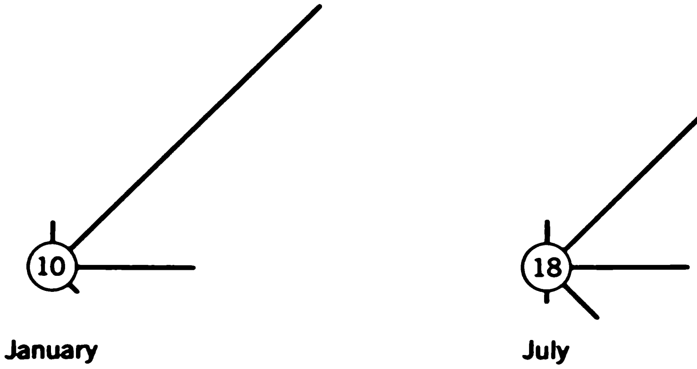


Figure 7.6
Wind roses for Georgetown, Guyana.

well defined, winter weather is controlled to a large extent by air moving off the land and hence is somewhat drier. There may be brief local reversals of the typical winter outflow and accompanying showers. Where the winter is not dominated by a strong offshore monsoon, the mere lessening of onshore movement of warm, moist air results in lower monthly precipitation averages. (See Figure 7.6.)

Some stations which show the general characteristics of the monsoon tropical climate depart from the rigid pattern of "wet summers and dry winters." The northeast coast of South America, for example, has its heaviest rains in December and January, when onshore trades are strongest, and in May through July, when dominated by the intertropical convergence, but the August-through-October period is relatively dry. Cayenne, French Guiana, with precipitation maximums in January and May, illustrates these features.

WET-AND-DRY TROPICS

As the name implies, the wet-and-dry tropical climate has alternating wet and dry seasons. It is transitional between the rainy and monsoon tropics on the one hand and the tropical arid and semiarid climates on the other. There is a distinctly dry period of two to four months, usually coinciding with the winter, and annual rainfall totals are less than in the rainy tropics. Mean monthly temperatures range from 18°C to above 25°C. The wet-and-dry tropics differ from the monsoon tropics most significantly in the effect of the dry season on vegetation and crops.

The climate occurs between the latitudinal limits of approximately 5° to 10° on the equatorial side to 15° to 20° on the poleward, that is, between the average locations of the equatorial low and the subtropic highs. The principal areas having this climate are in western Central America, northwestern South America, the interior uplands of Brazil and adjacent Bolivia and Paraguay, south-central and eastern Africa, western Madagascar, parts of India and Southeast Asia, and northern Australia.

Owing to a net excess of insolation over outgoing radiation the wet-and-dry tropics have generally high temperatures throughout the year, although on the higher plateaus of South America and eastern Africa altitude produces lower averages. As in the monsoon tropics, there is a tendency toward a maximum temperature in late spring or early summer just before the season of greatest cloudiness and heavy rain. There may be a secondary maximum again just after the rainy season. Along the poleward margins annual ranges are greatest, and, in general, the annual ranges are greater than in the rany tropics.

Diurnal ranges of temperature are greatest in the dry season and at higher elevations and are somewhat greater than in the rainy tropics. Nighttime temperatures may drop below 15°C in winter. The daytime temperatures of 25° to 30°C in winter are offset to some extent by the lower relative humidity. In summer, on the other hand, the diurnal range is small, temperatures are high, and these factors, combined with the rain and high humidity, produce oppressive conditions similar to the rainy tropics. Relief from the heat and humidity can be found only along coasts with strong sea breezes or at high altitudes.

The outstanding feature of precipitation (indeed of the climate) in the wet-and-dry tropics is the marked seasonal contrast. Many stations have one or more months with no precipitation recorded over a period of several years, whereas the wettest month has an average of 25 cm or more. Annual totals generally range from 100 to 150 cm, appreciably lower than in the rainy tropics. On the margins of the climatic region nearest the equator, the dry season is short and it is difficult to define the boundary between the wet-and-dry type and the monsoon tropics or the rainy tropics. Along the poleward margins the dry season is prolonged and conditions grade into the tropical semiarid climate, where potential evapotranspiration exceeds precipitation even in the wet season. Normally winter is the dry season, but along the Coromandel Coast of India spring is the dry period. Madras has a precipitation maximum in late autumn and early winter, when the monsoon circulation brings air across the warm Bay of Bengal from the northeast. The accompanying table shows climatic data for Madras, India, and Cuiaba, Brazil, both wet-and-dry tropics.

Madras, India 13°N; 16 m

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	24	26	28	30	33	32	31	30	30	28	26	25	29
P (mm)	24	7	15	25	52	53	83	124	118	267	308	157	1,233

Cuiaba, Brazil 16°S; 165 m

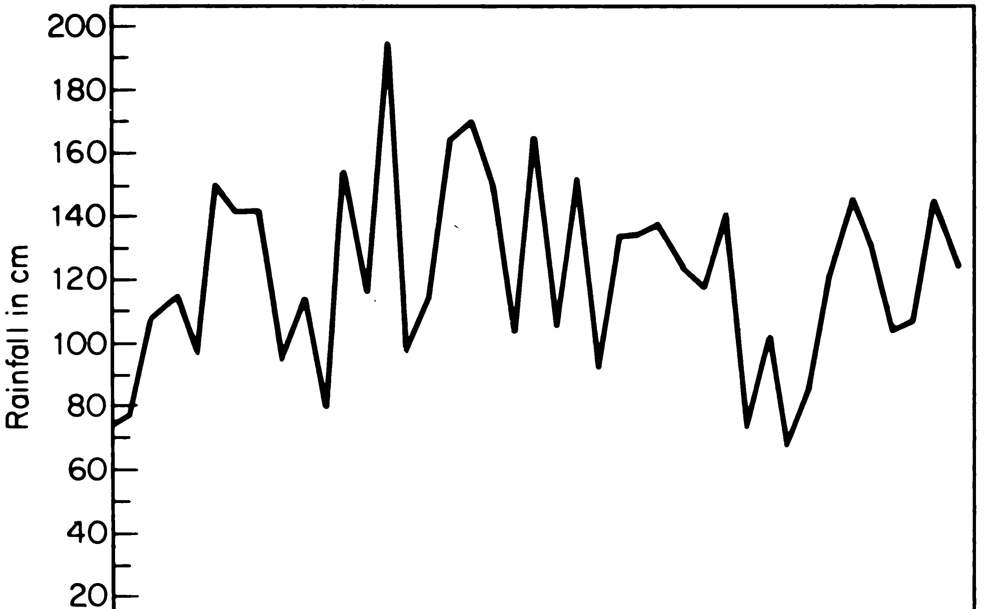
T (°C)	26	26	26	26	24	23	23	25	27	27	27	27	26
P (mm)	216	198	232	116	52	13	9	12	37	130	165	195	1,375

Reliability of precipitation is less than in the rainy tropics. A year with destructive floods may also have drought with serious economic consequences. Unseasonable rain or drought can be as disastrous as abnormal rainfall totals. The graph of annual rainfall over a 40-year period at Nagpur, India, illustrates the variability from year to year (Figure 7.7).

Precipitation in the wet-and-dry tropics is associated with thunderstorms and weak tropical lows. Thundershower activity is greatest at the beginning and end of the rainy season, when violent storms are interspersed with sunny periods. As the rainy season becomes established, shallow tropical lows of the type found in the rainy tropics bring long periods of rain and cloudiness. The dry season weather is essentially like that of the tropical deserts, with only erratic showers.

Although tropical air masses exercise the principal control over the wet-and-dry tropical climate, a marked seasonal alternation of dominant types is evident. In the wet season the equatorial trough and equatorial air masses control the climate, bringing weather not far different

Figure 7.7
Variations in annual rainfall at Nagpur, India. (Data from Government of India Meteorological Department)



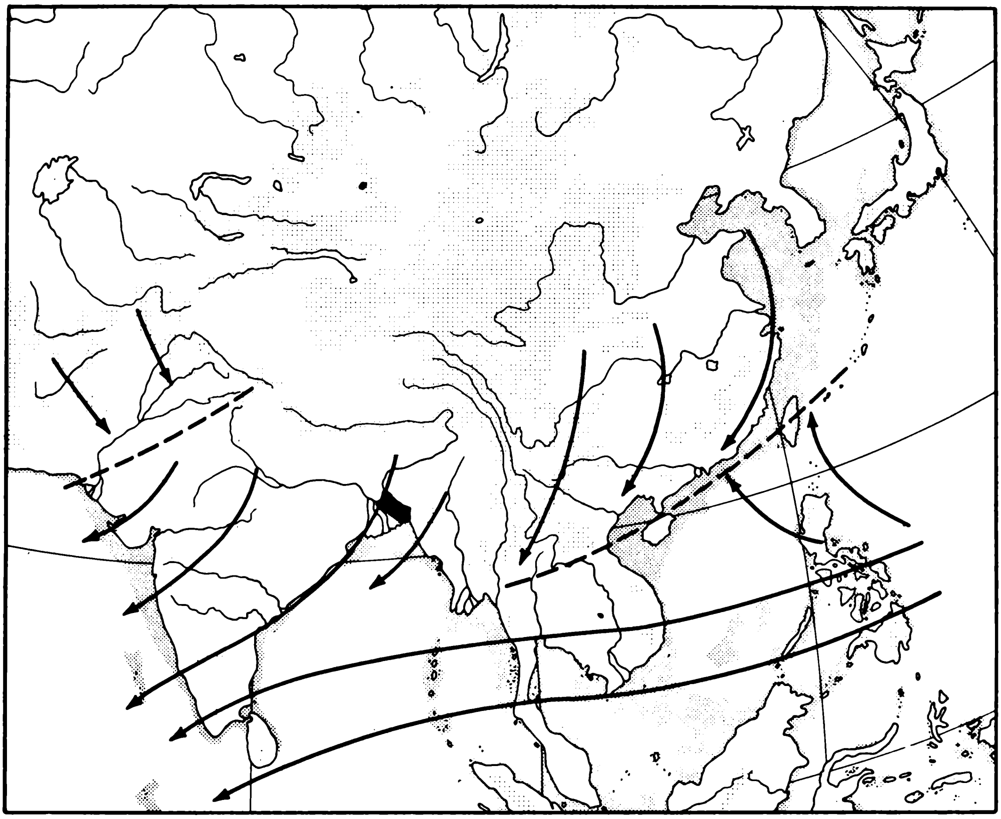
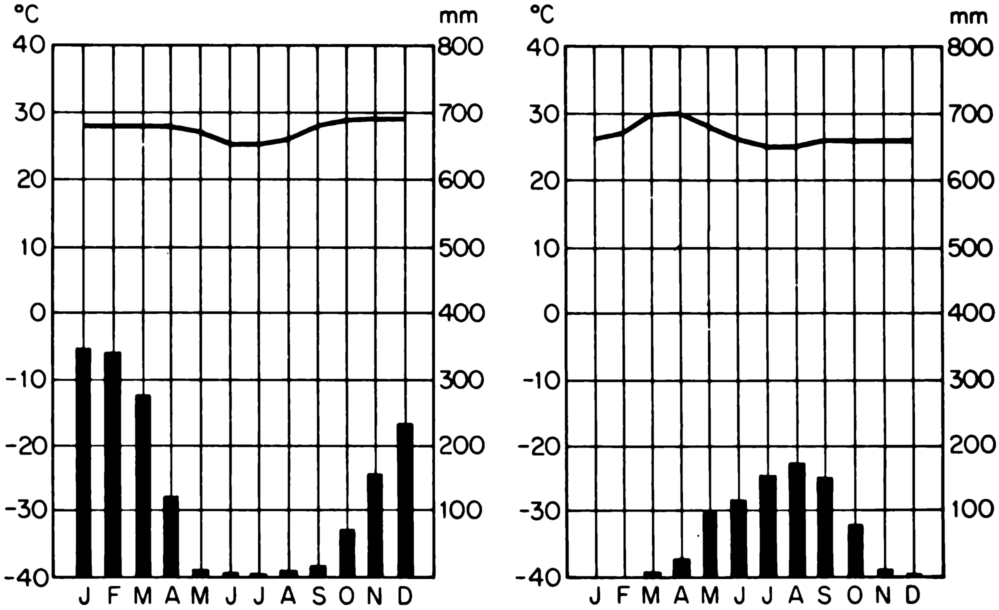


Figure 7.8
Mean air trajectories across southern Asia in January. (After Watts, Thompson, and others)

from the rainy tropics. This results directly from the poleward migration of the wind and pressure belts of the primary circulation in summer, when the sun is at or near the solstice. Although the effects of the sun's apparent poleward migration lag behind the noon sun's latitudinal position by a month or six weeks, they are nevertheless associated with the summer half of the year and indirectly produce rainy weather which gradually moves poleward until after the reversal of the sun's migration at the tropic. Even in India and parts of southeastern Asia, where the monsoon circulation is strong, the equatorial trough lies north of the equator in summer and plays a complementary role to the monsoon in accounting for summer rains. In winter the wet-and-dry tropics are dominated by the subsiding air masses of the subtropic highs, which have gradually moved toward the equator, bringing arid conditions (Figure 7.8). In "monsoon Asia" and in northern Australia this effect is again complementary to the monsoon, which in winter blows off the continent. Thus the climatic region is under the influence of unstable maritime air through most of the rainy season and of stable continental air in the dry season. (See Figure 7.9.)



Darwin, Australia	12°S; 27m
Mean annual temperature	28°C
Annual temperature range	4°
Mean annual precipitation	1562mm

Malakal, Sudan	10°N; 389m
Mean annual temperature	27°C
Annual temperature range	5°
Mean annual precipitation	783mm

Figure 7.9
Climatic graphs for Malakal and Darwin, wet-and-dry tropics.

TROPICAL ARID AND SEMIARID CLIMATES

The arid and semiarid climates of tropical latitudes have numerous features in common, their differences being more in degree than in kind. The semiarid types are essentially a transition zone from the very dry regions to the bordering moister climates. The distinctive characteristic of these climates is the lack of sufficient rainfall to sustain dense vegetative growth. They are differentiated from the mid-latitude counterparts by their higher average temperatures and the consequent need for more rainfall to offset evaporation and transpiration.

The tropical arid and semiarid climates are centered approximately on the latitudes 20° to 25° north and south, where the prevailing air masses are those that subside in the subtropic highs. The process of subsidence results in adiabatic heating and low relative humidity and thus dries the land surfaces. Even where surface heating produces intense low pressure the convection layer is shallow because of upper-air inversions whose effects extend the arid conditions well into the latitudes of the trades. The trade-wind inversion (a subsidence inversion) tends to restrict vertical cloud development. The chief areas of tropical arid (or desert) climate are in northwestern Mexico and southwestern United States, along the west coast of Peru and northern Chile, the Sahara and Somalia in North Africa, portions of southwestern Asia from Arabia to Pakistan, the west coast of southern Africa, and central Australia. Extensive oceanic areas that experience the effects of air mass subsidence in the subtropic highs also have sparse rainfall. Sal in the Cape Verde Islands and Ascension Island in the South Atlantic have mean annual totals of only 10.8 and 12.2 cm, respectively. With certain exceptions, notably along west coasts, the core areas of arid climates have transitional semiarid belts situated more or less concentrically around the wetter margins. On the Deccan Plateau of India an elongated area of semiarid climate occurs without a contiguous area of true arid climate. Where tropical arid climates are located along west coasts of continents, ocean currents modify desert conditions. The four principal instances are as follows:

Region	*Ocean Current*	*Desert Name*
Lower California and Sonora	California	Sonoran
Coastal Peru and Chile	Humboldt or Peru	Peru and Atacama
Northwest African Coast	Canaries	Sahara
Southwest Africa	Benguela	Namib

Upwelling associated with these currents produces cool temperatures along the immediate coasts and increases stability in the lower air layers, thereby reducing the tendency to cloud formation and precipitation. Advection fogs, stratus, and even drizzle are common along the coasts thus affected, however. In combination with lower temperatures they reduce the water need from precipitation.

Comparison of the climatic data for Walvis Bay, a coastal station of Southwest Africa, and Windhoek, inland but at an elevation of 1,728 m, reveals the influence of the cold Benguela Current (see the accompanying table). In view of its altitude, Windhoek might be expected to have cooler summer temperatures. Instead, it is both warmer and wetter than Walvis Bay.

Walvis Bay, Southwest Africa 23°S; 7 m

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	19	19	19	18	17	16	14	14	14	15	17	18	17
P (mm)	2	5	8	3	3	0	1	3	1	1	2	1	30

Windhoek, Southwest Africa 23°S; 1,728 m

T (°C)	23	21	21	19	16	13	13	16	20	22	23	23	19
P (mm)	77	73	81	38	6	1	1	0	1	12	38	47	370

Another influence upon arid and semiarid climates of the tropics is the mountain barrier effect. There is no arid area which can be ascribed primarily to mountain barriers, although the Andes in South America and the Atlas Range in northwestern Africa form rather definite rainfall boundaries. A more important effect of mountains in these climates is to induce slightly greater rainfall on their slopes. Examples are to be found in the Ahaggar Mountains in the Sahara, the highlands of Yemen, and along the ranges of Iran and Afghanistan. Not only is the precipitation greater along the mountain slopes, but the temperatures are lower owing to altitude, and potential evapotranspiration is therefore less.

The arid and semiarid areas of East Africa and eastern Brazil must be regarded as atypical locations of these climates in view of their position in latitudes frequently visited by the intertropical convergence. In parts of northern Kenya average annual rainfall is less than 15 cm. This is presumed to result from subsidence, divergence, and consequent stability of air that reaches the area, especially at upper levels, from the Indian Ocean. In far eastern Brazil the stable air of an offshore anticyclone produces a dry summer. In winter the moist southeast trades which bring rainfall to the coast south of Cape São Roque do not penetrate strongly beyond the uplands in most years.

Maximum temperatures in the tropical arid climate are the highest in the world. Extreme maximums above 50°C are common. Azizia, Libya, recorded an official air temperature of 58°C (136°F) on September 13, 1922. Ground temperatures are much higher under intense insolation. Monthly means of the summer months exceed 30°C at many tropical desert locations. El Golea and Jacobabad are example stations (see accompanying table) that have high summer temperatures, contrasting sharply with Walvis Bay and Windhoek.

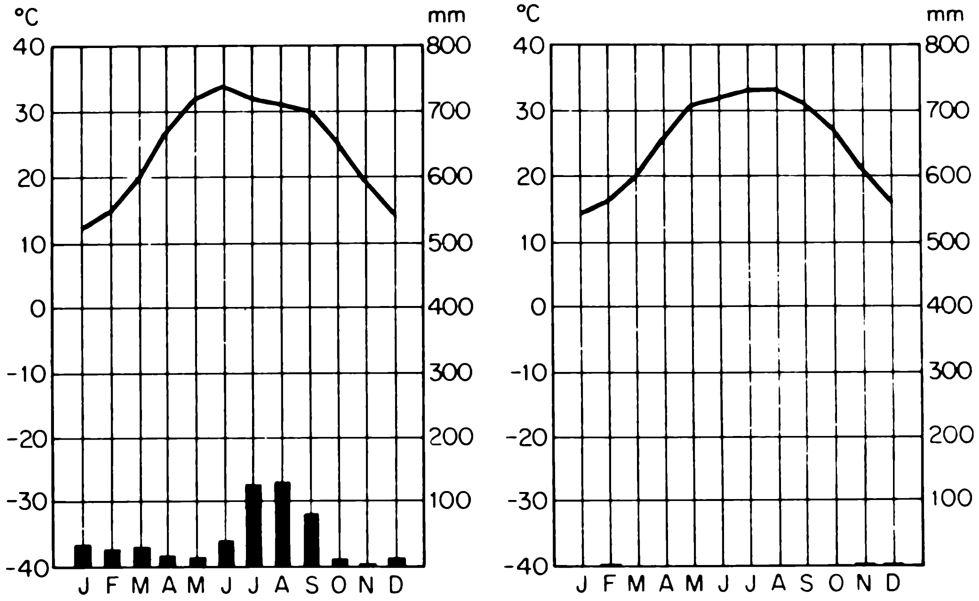
El Golea, Algeria 31°N; 398 m

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	9	12	16	21	26	32	34	33	30	22	15	10	22
P (mm)	8	1	6	4	T	1	0	T	2	4	4	14	44

Jacobabad, Pakistan 28°N; 56 m

T (°C)	15	18	24	30	35	37	35	34	32	28	22	17	27
P (mm	8	8	7	2	4	6	37	22	1	0	1	3	99

Annual range of temperature is far greater in the tropical deserts than in the rainy tropics. This is due in part to the generally higher latitudes of the former but results primarily from the clear skies that permit high net radiation in summer and a reduced radiation budget in winter. The annual ranges for El Golea and Jacobabad are 25C° and 22C°, respectively. Compare these with rainy tropical stations, for ex-



Lahore, Pakistan	32°N;214m
Mean annual temperature	24°C
Annual temperature range	22°
Mean annual precipitation	492mm

Luxor, Egypt	26°N;95m
Mean annual temperature	25°C
Annual temperature range	19°
Mean annual precipitation	1mm

Figure 7.10
Climatic graphs for Lahore and Luxor, tropical arid.

ample, Coquilhatville or Singapore (Figure 7.1). Coastal stations have smaller annual ranges. Just as the annual range increases over dry lands and under clear skies, so is the diurnal range greater. Diurnal ranges of 15° to 25C° are common. On winter nights when long-wave radiational losses are rapid temperatures occasionally drop below freezing. Nocturnal frosts are particularly trying because of the contrast with the warm days.

Temperatures in the semiarid tropical climate are similar to those of the arid type. (See climatic graphs for Lahore and Luxor in Figure 7.10.) As would be expected, the averages for semiarid climates on the high-latitude borders of the deserts are generally lower and the ranges greater than on the savanna margins nearer the equator.

Precipitation in the tropical arid climates is not only low in amount but it is also erratic. There is no definite seasonal regime that could be termed characteristic of tropical arid stations; the climate is identified by lack of effective precipitation rather than a unique seasonal distribution. For example, stations near the margin of the monsoon tropics have a monsoonal regime, whereas those on west coasts bordering the dry summer subtropics have a winter maximum. The stations with no significant precipitation in any month can hardly be said to have a regime of precipitation. Arica, in northern Chile, has the record minimum average annual rainfall—only 0.5 mm over a 43-year period. Iquique, on the coast south of Arica, had no rain for 14 years, but has a statistical average of about 15 mm annually. In the semiarid climate the seasonal distribution becomes better defined as a rule and is more important in its effect on vegetation and land use. On the other hand, the variability of precipitation is a more critical factor in the semiarid climate than in the true desert. In wetter-than-average years settlers may venture into the semiarid grasslands only to find that the succeeding years are too dry for their type of land use.

Most of the precipitation of the tropical deserts is of the thunderstorm type. It comes in downpours that rapidly exceed the absorptive capacity of the soil, and because there is little vegetative cover the runoff is great, often of flood proportions. Thus a relatively small part of the rain is retained as soil moisture. In the semiarid climates there may be longer periods of rain associated with weak lows. In the poleward locations of the semiarid climate, winter rain comes with the mid-latitude cyclones which occasionally reach into these areas. On the margins toward the equator the intertropical convergence may bring limited summer rain.

The fogs and lower temperatures associated with the cold currents and upwelling along the coastal deserts modify the moisture conditions sufficiently to permit growth of some low forms of vegetation in spite of almost complete lack of rainfall. On the coast of Peru fog is so heavy at times that it is virtually a drizzle and is locally referred to as the *garua*. It is along the Peruvian coast that an interesting and unusual feature of variability in desert precipitation occurs. When the subtropic high moves abnormally seaward and the Peru Current flows at some distance from the coast, warm water from the north, known as *El Niño,* spreads out over the cold water along the shore. If the intertropical convergence has moved southward, as may happen in the Southern Hemisphere summer, there may be heavy rains typical of the rainy tropics. The combination of conditions necessary to produce this phenomenon ordinarily prevails for only a short period. One of the worst occasions was in 1925, when ruins of sun-dried bricks that had stood for centuries in the vicinity of Trujillo were nearly washed away.

Additional Readings

BARRETT, ERIC C., "The Tropical Far East: ESSA Satellite Evaluations of High Season Climatic Patterns," *Geog. Journ.,* 137 (1971), 535–55.

CHANG, JEN-HU, "Comparative Climatology of the Tropical Western Margins of the Northern Oceans," *Ann. Assoc. Am. Geogrs.*, 52, 2 (1962), 221–27.

DAVIES, H. R. J., "Khartoum Weather," *Journ. of Geog.*, 58, 6 (1959), 286–91.

FRISBY, E. M., AND H. W. SANSOM, "Hail Incidence in the Tropics," *Journ. Appl. Met.*, 6, 2 (1967), 339–54.

FOSBERG, F. R., B. J. GARNIER, AND A. W. KÜCHLER, "Delimitation of the Humid Tropics," *Geog. Rev.*, 51, 3 (1961), 333–47.

GRAY, WILLIAM M., "Global View of the Origin of Tropical Disturbances and Storms," *Mon. Wea. Rev.*, 96, 10 (1968), 669–700.

GRIFFITHS, JOHN F., ed., *Climates of Africa,* Vol. 10, World Survey of Climatology. New York: American Elsevier Publishing Company, 1971.

LOCKWOOD, J. G., "Synoptic Disturbances near the Equator," *Weather,* 20, 9 (1965), 279–82, 295.

NIEUWOLT, S., "Diurnal Rainfall Variation in Malaya," *Ann. Assoc. Am. Geogrs.*, 58, 2 (1968), 313–26.

PORTIG, W. H., "Central American Rainfall," *Geog. Rev.*, 55, 1 (1965), 68–90.

RAMAGE, C. S., *Monsoon Meteorology*. New York: Academic Press Inc., 1971.

SELLERS, WILLIAM D., "Potential Evapotranspiration in Arid Regions," *Journ. Appl. Met.*, 3 (1964), 96–104.

chapter eight

Climates Dominated by Tropical and Polar Air Masses

In the middle latitudes of both hemispheres lie the battlegrounds of contrasting air masses, where warm tropical air meets cold polar air along the ever-fluctuating polar fronts. Seasons are primarily warm and cold rather than wet and dry as in the tropics; the changing temperatures from season to season and with the advance and retreat of air masses play a much larger part in influencing man's activities. Instead of the monotony of the tropical climates there are both periodic and erratic weather changes, making these latitudes "intemperate" rather than temperate as they are commonly designated. These are the belts of the extratropical cyclones, where most of the precipitation is associated with fronts. All the climates of this group are subject to frost and all experience snow, although the amount and duration of snow cover vary widely. The principal air masses that dominate this group of climates are the cT, mT, cP, mP and their modifications. The interactions of these air masses produce the six major climatic types discussed in this chapter.

DRY SUMMER SUBTROPICS

The dry summer subtropical climate prevails on the west coasts of continents in the lower middle latitudes where the controlling air masses are mT out of the eastern margins of the subtropic highs. Only in the Mediterranean Basin does this climate extend far eastward from a major ocean. From this region the climate has taken another common name, the Mediterranean type. Other locations are central California, central Chile, the southern tip of Africa, southwestern Australia, and an area in the southeastern part of South Australia and adjacent Victoria.

The chief features of the climate are a hot, dry summer and a mild, rainier winter. During the summer the climate comes under the influence of stable air which flows out of the oceanic subtropic high cells to the west. (See Figure 8.1.) (In the case of the Mediterranean Basin the zone of subsidence in the subtropic high extends far into the eastern end of the Mediterranean Sea.) Under these conditions the climate is very much like that of the tropical arid and semiarid types. In winter the migration of the global circulation brings to these regions the tropical margins of the westerlies with their cyclonic passages and occasional invasions of modified polar air masses.

Annual temperature averages are somewhat lower than in the tropical climates; this is a *sub*tropical climate in terms of latitudinal location and temperature. Monthly averages in summer do not often exceed 27°C, although extreme maximums of over 38°C have been recorded at numerous stations. Coastal locations have much cooler summers because of a marine effect which is increased by cool ocean currents. An exception is found on Mediterranean shores, where the absence of cool currents and the generally warmer water temperatures cause higher summer temperatures than in other areas having this type of climate. A comparison of Sacramento and San Francisco affords an illustration of the influence of the ocean (see the accompanying table). Sacramento has a July mean of 25°C, whereas San Francisco on the coast has only 17°C. Note that San Francisco does not achieve its maxi-

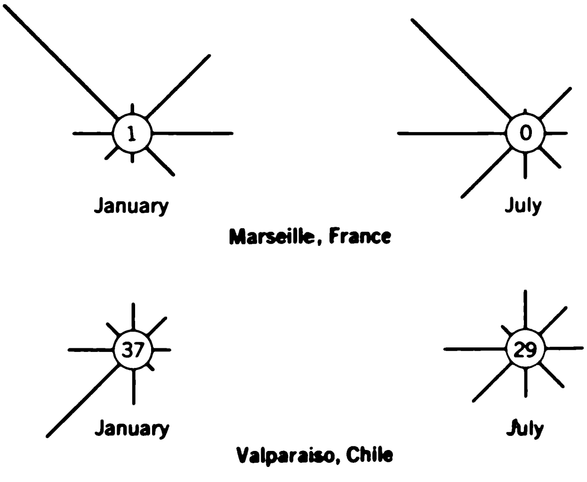


Figure 8.1
Wind roses for Marseille and Valparaiso.

mum mean monthly temperature until September, when the coastal waters are warmer and there is less fog.

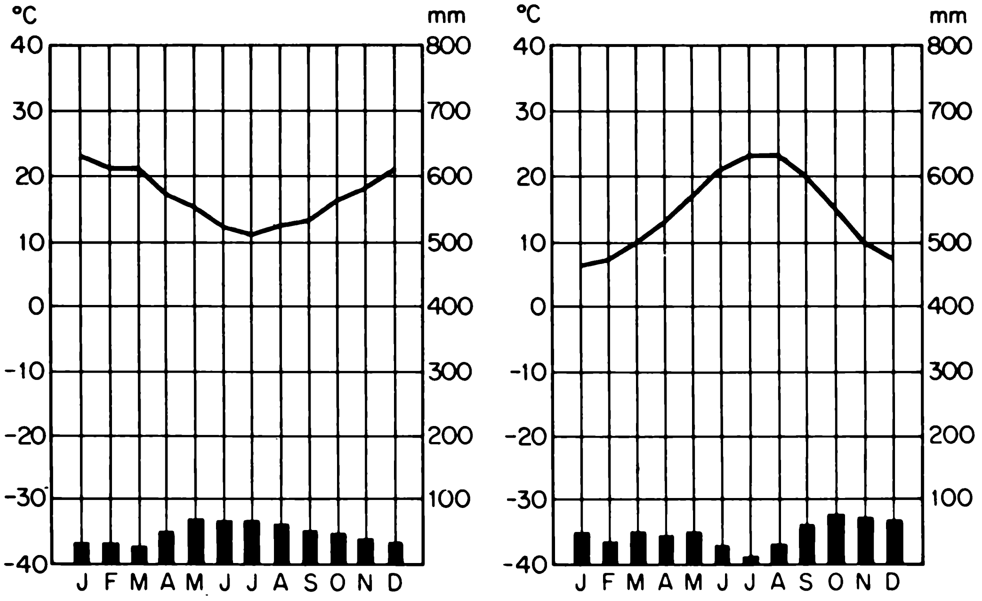
Sacramento, California 39°N; 13 m

	J	F	M	A	M	J	J	A	S	O	N	D	Yr
T (°)	8	10	12	16	19	22	25	24	23	18	12	9	17
P (mm)	81	76	60	36	15	3	T	1	5	20	37	82	414
San Francisco, California 38°N; 27 m													
T (°C)	9	10	12	13	15	16	17	17	18	16	13	10	14
P (mm)	102	88	68	33	12	3	T	1	5	19	40	104	475

The summer temperatures of the interior locations are reminiscent of the deserts. Daily maximums are high, frequently above 30°C, but the nighttime low may be below 15°C. With clear skies and low relative humidity, daytime heating is intense, but nocturnal cooling is rapid. Along the coasts, however, diurnal temperature ranges are much lower. The mean daily maximum for January at Valparaiso on the coast of Chile is about 21°C, whereas at Santiago in the interior it is 30°C. Yet Santiago has a lower mean daily minimum in the same month. Winter is a distinctly cooler season in the dry summer subtropics, the coolest month having a mean temperature usually below 10°C. As in summer, the diurnal ranges are greatest in the interior. Winter frosts occur but they are rarely severe. On the few occasions when night temperatures drop well below the freezing point there is great damage to citrus and other crops. Freezing temperatures in the dry summer subtropics are ordinarily the result of rapid nocturnal radiation and air drainage in the lower layers of a polar continental air mass that has invaded the subtropical latitudes, but sometimes freezing temperatures prevail throughout such air masses.

Katabatic, or gravity, winds sometimes plague the coastal regions in winter. The *mistral* in southern France and the *bora* along the Adriatic coast of Yugoslavia flow seaward from interior plateaus to create unpleasantly cool, dry conditions. In Southern California a hot, gusty wind known as the *Santa Ana* blows toward the coast in winter when a high pressure center is developed over the western United States. Wind speeds sometimes reach 45 knots, and because of low relative humidity, the air becomes dusty and forest fire danger is often critical. In South Africa the hot, dry *berg* blows from the plateau as cyclonic circulation replaces high pressure.

Annual precipitation totals in this climate generally fall within the range of 35 to 90 cm, the amount being least on the semiarid margins and increasing poleward toward the marine climate. The summers have little or no rain, and this feature combined with the high temperatures results in extremely low soil moisture content. (See Figure 8.2.) The meager summer rainfall consists largely of scattered showers attended by rapid runoff. Typical summer weather is an unbroken series of hot, sunny days. Along the immediate coasts, the modified temperatures are accompanied by a tendency to fogginess, especially where the effect of cold ocean currents is extended poleward from the desert coasts. Fog and dew are important factors in plant growth along these coasts. Winter precipitation comes from cyclonic storms, the moisture sources, for the most part, being maritime tropical and maritime polar air masses. Frontal rainfall associated with cyclones is frequently light and scattered because the low centers normally pass on the high-latitude side of the climatic regions. Occasionally a low passes directly over the region, bringing frontal conditions approximating those which are regular occurrences in the marine west-coast climate, but with rainfall concentrated in fewer rainy days. Stations with little or no rain in the driest summer month commonly have 7 to 10 cm in the wettest winter month. Snow is rare at lower elevations. Both snow and total precipitation increase on the slopes of mountains, which at their greater heights are properly classified with the highland climates.



Adelaide, Australia	35°S; 43m
Mean annual temperature	17°C
Annual temperature range	12°
Mean annual precipitation	523mm

Marseille, France	43°N; 3m
Mean annual temperature	14°C
Annual temperature range	17°
Mean annual precipitation	546mm

Figure 8.2
Climatic graphs for Adelaide and Marseille, dry summer subtropics.

HUMID SUBTROPICS

The humid subtropics lie in approximately the same latitudes as the dry summer subtropics, but they are on the eastern sides of the continents, where they come under the influence of unstable (or conditionally unstable) tropical air out of the western margins of the subtropic highs in summer. Annual precipitation is greater than in the dry summer subtropics, but in summer amounts may be inadequate to meet the needs of potential evapotranspiration. (See Figure 6.1, a water budget graph for Memphis.)

World distribution of the humid subtropical climate is chiefly in the following areas: southeastern United States, a belt along northern India, southern and eastern China, Taiwan and southern Japan, northeastern Argentina and adjacent areas, the Natal Coast of South Africa, and the eastern coast of Australia. A smaller area is found at the eastern end of the Black Sea. Note that although the poleward boundaries of these climates are essentially at the same latitude as the dry subtropics, they extend farther toward the equator and merge into one of the humid tropical types.

During the summer, the humid subtropics are dominated by maritime tropical air which flows inland from the anticyclonic circulation of the subtropic high cells. Summer temperatures and humidity are uncomfortably like those of the rainy tropics, for the air from the subtropic highs flows onto the land from the lower latitudes with a path over warm water. This is in contrast to the dry summer subtropics, where subsiding air moves out of the high cells from higher latitudes over colder water. Thus, the prevailing air masses over the dry summer subtropics are relatively stable and dry, whereas those of the humid subtropics are moist, warm, and unstable. (See Figure 8.3.) The prevalence of warm currents along east coasts in the subtropics and cold currents along the west coasts accentuates the contrast between the two climatic types. In winter, the humid subtropics are influenced primarily by the belt of mid-latitude cyclones, with the two main air mass types being polar continental and maritime tropical. In Asia, the monsoon circulation is superimposed upon the global circulation and winter months have less precipitation than summer owing to the prevailing flow of stable air off the continent.

Temperatures in the humid subtropics are similar to those in the dry subtropics, but, because of the high relative humidity and warmer water offshore, the summers are more like those in the wet tropical climates. The mean monthly temperature of the warmest month is around 27°C for most stations. Mean daily maximums reach 30° to 38°C, and extremes exceed 38°C. Diurnal ranges are small, the nights being

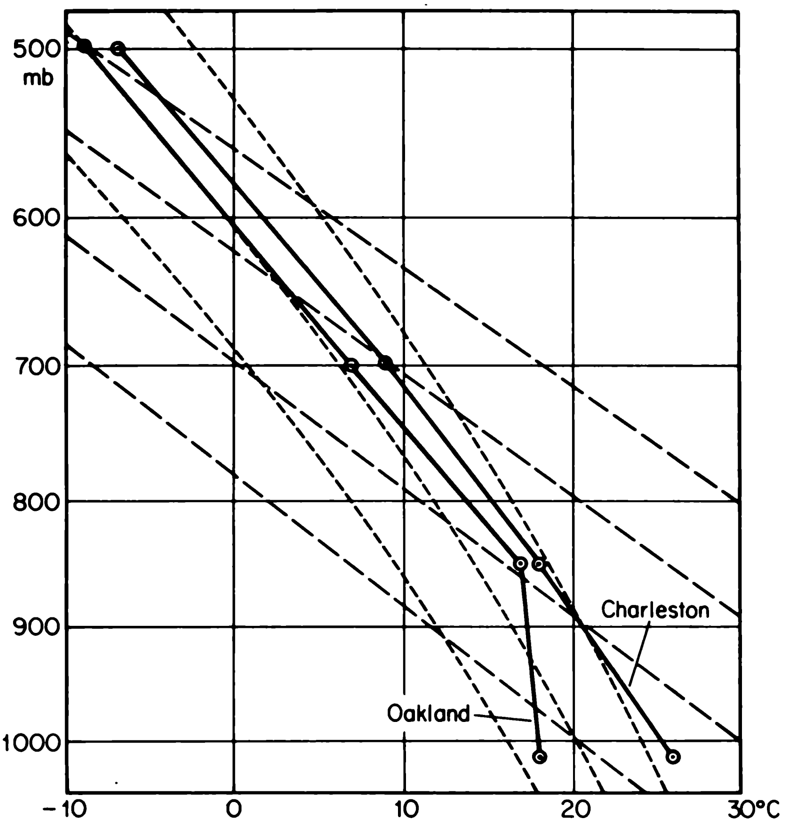


Figure 8.3
Mean vertical temperature soundings for July at Oakland, California, and Charleston, South Carolina. Note the greater stability of the surface layer at Oakland.

much more oppressive than in the dry summer subtropics. Furthermore, temperature variations from day to day are not very great. The growing season is long, but relief from summer conditions of heat and high relative humidity comes with the autumn as the first salients of polar air reach into these latitudes. Then the humid subtropics suddenly lose their similarity to the rainy tropics. Temperatures of the cold month average between 5° and 12°C for most areas in the climatic type. The major exceptions are in monsoon Asia. Hong Kong, has, as its lowest monthly mean, 15°C in January. Allahabad, India, has a January average of 16°C. These stations would be classified with the tropical climates but for their distinctly cooler winter seasons. The greater proportion of water in the Southern Hemisphere causes annual ranges to be generally lower than in the Northern Hemisphere in this climate. In both cases coastal stations show smaller annual ranges than those in the interior. Compare the temperature data for New Orleans with those for Buenos Aires and Memphis (see the accompanying table). Memphis, an interior station, has an annual range of 22C° as compared with 15C° at New Orleans, whereas Buenos Aires has a lower summer maximum and a 14C° annual range.

Buenos Aires, Argentina 35°S; 55m

	J	F	M	A	M	J	J	A	S	O	N	D	Yr
T (°C)	24	23	21	17	14	11	10	12	14	16	20	22	17
P (mm)	104	82	122	90	79	68	61	68	80	100	90	83	1,027

Memphis, Tennessee 35°N; 86 m

T (°C)	6	7	11	17	22	26	28	27	24	18	11	7	17
P (mm)	148	116	124	117	102	92	87	67	71	73	106		1,223

New Orleans, Louisiana 30°N; 17 m

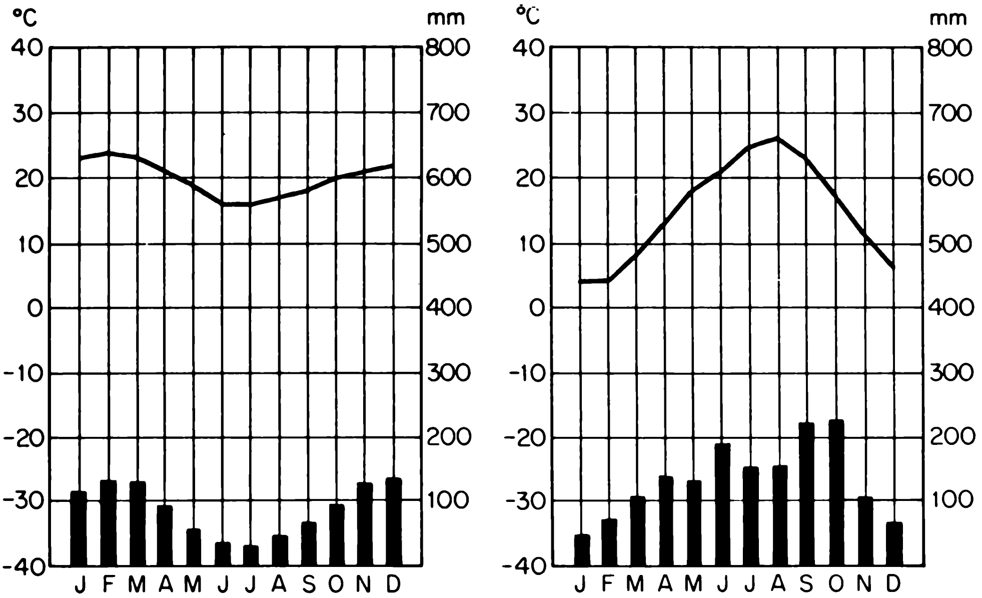
T (°C)	12	13	16	19	23	26	27	27	25	21	15	13	20
P (mm)	98	101	136	116	111	113	171	136	128	72	85	104	1,369

Day-to-day variation in temperature becomes much greater as polar and tropical air masses alternately advance and retreat in their battle for supremacy in the succession of cyclones. While the winters can best be described as mild, frosts are normal in the higher latitudes and occasionally plague the tropical margins. Extreme minimums of −6° to −12°C have occurred along the United States Gulf Coast. Severe frosts are a hazard to crops and orchards in Florida just as they are in California.

Typical annual averages of precipitation in the humid subtropics vary from 75 to 150 cm. The lower averages are found in the regions bordering the semiarid climates, the greater amounts along mountain ranges and in the tropical extensions of the climatic type. Most areas have a fairly uniform distribution of rainfall throughout the year; 8 to 15 cm are typical monthly averages. Nevertheless, great variety can be found in precipitation regimes. In the southeastern United States the primary monthly maximum occurs in March at Vicksburg, in May at Little Rock, and in July at Mobile. Miami has a maximum in September owing to hurricane activity.

In India and southeastern Asia the monsoon produces a distinctly dry winter as dry continental air flows toward the coasts. Tokyo's autumn maximum represents the blending of a monsoon effect with tropical cyclonic storms. (See Figure 8.4.) Its secondary maximum in early summer results from a southwesterly stream of warm, moist, unstable air of equatorial origin. The rains of this season are known as the *Baiu* in Japan.

For the climate as a whole, summer precipitation is primarily from air mass thunderstorms. (See Figure 8.5.) Hurricanes or typhoons (or their remnants) visit the coasts occasionally in late summer and autumn

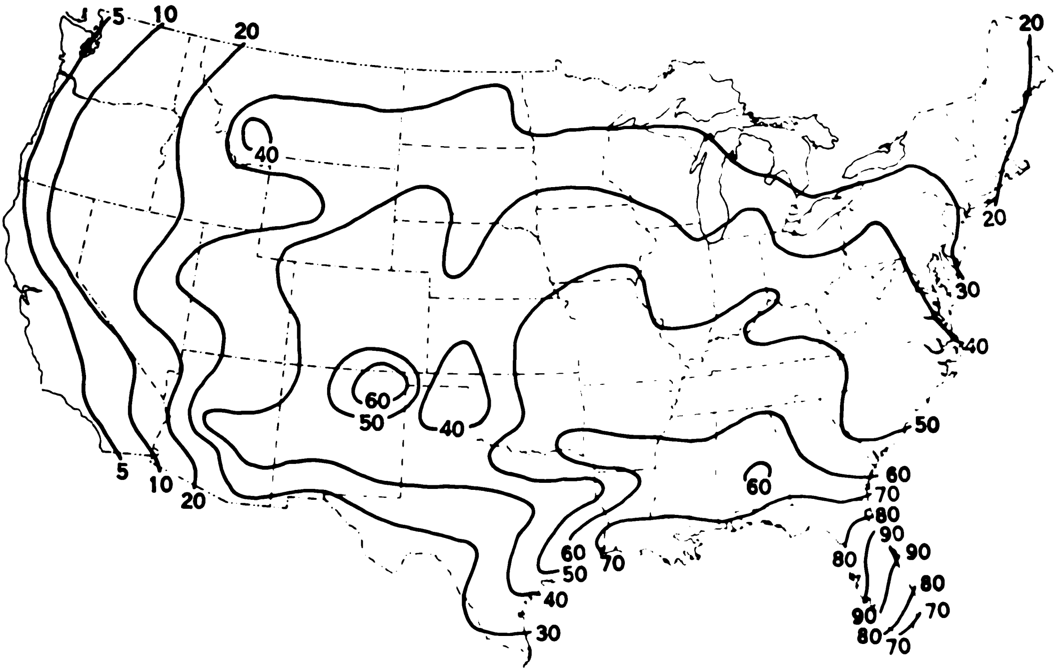


Durban, South Africa	30°S; 8m
Mean annual temperature	20°C
Annual temperature range	8°
Mean annual precipitation	1044 mm

Tokyo, Japan	36°N; 6m
Mean annual temperature	15°C
Annual temperature range	22°
Mean annual precipitation	1563 mm

Figure 8.4
Climatic graphs for Durban and Tokyo, humid subtropics.

Figure 8.5
Mean annual number of days with thunderstorms in the contiguous United States. (NOAA)



and increase the monthly rainfall totals. As the polar front moves into these regions, frontal precipitation and thunderstorms of the frontal type are common in the converging poleward flow of air drawn into the warm sectors of cyclones.

Winter precipitation is mostly related to fronts and sometimes comes in the form of snow. Rainfall is lighter and more continuous; rainy days are greater in number, and cool, overcast days are more frequent. Frontal and ground fogs both occur in the humid subtropics, primarily in winter. Coastal fogs of the advection type are generally lacking because of the absence of cold currents in the adjacent oceans.

Mountain barriers have a significant effect on the distribution of precipitation in the humid subtropics just as in every other climate. One of the most unusual climatic stations in the world is Cherrapunji, India, to which allusion has already been made in Chapter 3 (see also the accompanying table). Cherrapunji is located at 1,313 m above sea level on the south side of the Khasi Hills in northeastern India. It is more than 300 km from the Bay of Bengal, but when the summer monsoon blows strongly the large area of lowlands allows moist air to pass inland, where it is forced upward in a funnel-shaped depression. Flood waters in the lowlands are warmer than the sea and probably contribute a share of the moisture that reaches the uplands. The dividing effect of the Khasi Hills acting as a mountain barrier is shown by the fact that Shillong, at 1,500 m but on the northern slope 40 km away, had only 241 cm mean annual rainfall in the period 1931–60. Were it not for the lower monthly temperatures and the relatively high annual range of temperature, Cherrapunji would be classified as a monsoon tropical station.

Cherrapunji, India 25°N; 1,313 m

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	12	13	17	19	19	20	20	20	20	19	16	13	17
P (mm)	20	41	179	605	1,705	2,875	2,455	1,827	1,231	447	47	5	11,437

MARINE CLIMATE

The characteristic location of the marine climate is on the west coasts of continents poleward from the dry summer subtropics, where air of oceanic origin flows onshore. The climate is frequently referred to as the marine west-coast type, although it is found in island areas and certain southeast coasts. The main locations are on the west coast of North America from California to southeastern Alaska; the British Isles; northwestern Europe from Portugal to the Scandinavian Peninsula; southern Chile; southeastern Australia; and New Zealand.

Most areas having this type of climate receive ample precipitation in winter, but in summer potential evapotranspiration exceeds rainfall. Temperatures are mild for the latitude in winter and cool in summer (see data for Portland and Melbourne in the accompanying table). The dominant air masses are of maritime polar origin, but modified mT and cP also affect the climate. In winter there is a succession of cyclones as the westerlies and the polar front prevail in these latitudes. There are fewer frontal passages in the marine climate in summer, when the subtropical oceanic high cells reach their highest latitudes, diverting cyclonic storms poleward. Along the eastern margin of the Pacific high an outflow of stable, subsiding air brings distinctly drier conditions to the North American Pacific Coast.

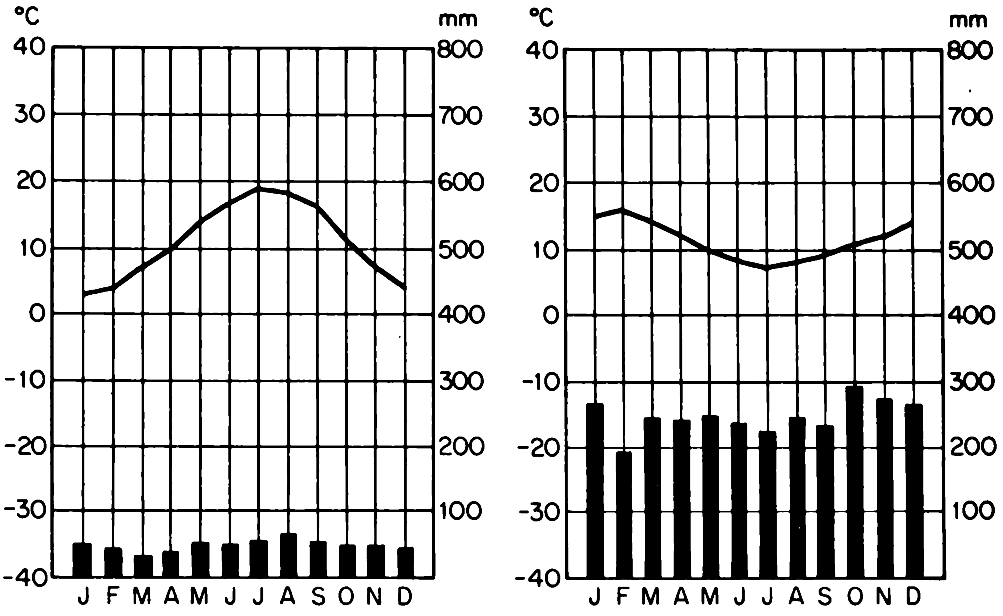
Portland, Oregon 46°N; 23 m

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	4	6	8	11	14	17	20	19	17	12	7	5	12
P (mm)	136	107	97	53	51	42	10	17	41	92	135	162	944

Melbourne, Australia 38°S; 44 m

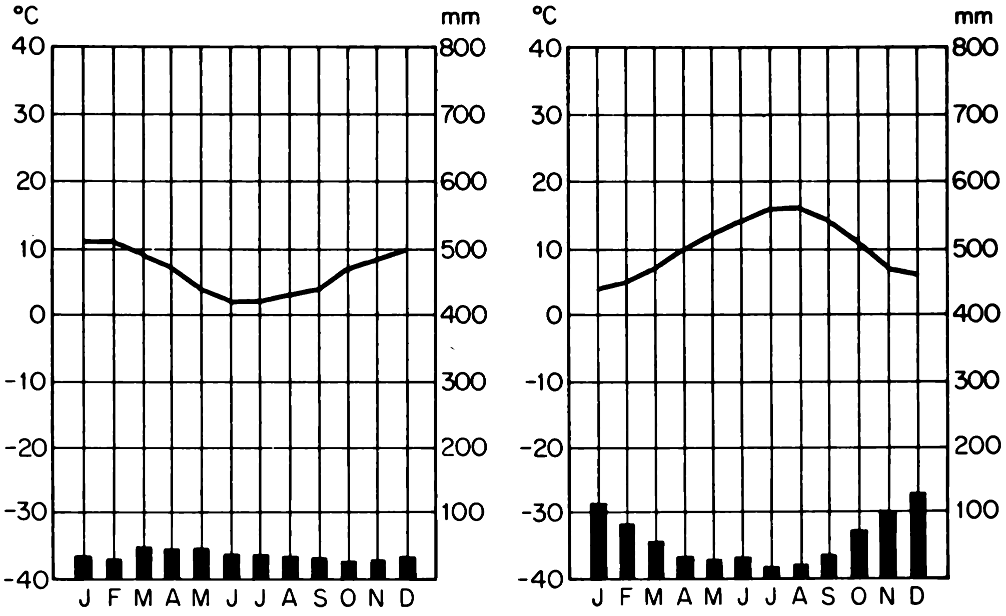
	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	20	20	18	15	12	10	10	10	12	14	16	18	15
P (mm)	45	59	50	69	54	52	54	50	58	74	70	58	691

Mean annual temperatures in the marine climate are mostly in the range of 7° to 13°C, unless modified by altitude. The mean monthly temperature of the warmest month is usually 15° to 20°C. Daily maximum temperatures do not often exceed 25°C, but extreme maximums of over 35°C sometimes occur when continental air temporarily invades a coastal area. Nights are cool, partly owing to nocturnal radiation and partly because of the generally low temperatures produced by winds off the sea. The marine influence is seen in low diurnal ranges as well as low annual ranges of temperature, but the effect decreases inland. Horizontal temperature gradients are greater from the coasts toward inland locations than in north-south directions, because the transport of heat from ocean sources is more significant than latitudinal variation of net radiation in this climate. Wintertime anomalies are even greater than in summer, temperatures being from 5° to 15C° warmer than the normal for the latitudes. Long periods with overcast skies and the prevailing onshore flow of maritime air in winter keep diurnal ranges small. Off northwest Europe and the British Isles, the Gulf Stream and the North Atlantic Drift carry the abnormally warm winter temperatures into high latitudes. Mean monthly temperatures of the coldest month are above freezing throughout the climatic region. Extreme minimums below zero occur at the higher latitudes and at inland and upland stations, but prolonged periods of continuous freezing weather are unusual at the



Paris, France	49°N; 53m
Mean annual temperature	11°C
Annual temperature range	16°
Mean annual precipitation	585mm

Hokitika, New Zealand	43°S; 3m
Mean annual temperature	11°C
Annual temperature range	9°
Mean annual precipitation	2907mm



Punta Arenas, Chile	53°S; 28m
Mean annual temperature	6°C
Annual temperature range	9°
Mean annual precipitation	407mm

Victoria, British Columbia	48°N; 69m
Mean annual temperature	10°C
Annual temperature range	12°
Mean annual precipitation	698mm

Figure 8.6
Climatic graphs for Paris, Hokitika, Punta Arenas, and Victoria, marine climates.

coasts. Outbreaks of polar continental air are responsible for periods of clear, freezing weather. At such times the wind is from an easterly direction, blowing out of a continental high center (chiefly cP or A air masses). The North American Pacific Coast is protected to some extent from the cold continental air by the mountain barriers of the Rockies and Cascades. Even when the air pushes to the coast, the stable-air foehn effect modifies its chilling effects appreciably. In northwest Europe, the marine climate extends farther inland than on any other continent, largely because of the absence of an effective mountain barrier. Not only do the maritime effects reach farther eastward, but the occasional polar outbreaks of northern Europe cover a much larger area, often including the coasts. Boundaries between marine and continental climates are sharply defined by mountains in North America and in southern Chile but are transition zones in Europe.

There is a wide range of annual precipitation in the marine climate, with variations from less than 50 to more than 250 cm. (See Figure 8.6.)

Although the marine climate as a whole must be characterized as having a deficiency of precipitation in summer, its dry season is neither so long nor so pronounced as that of the dry summer subtropics. As would be expected, the margins of the marine climate lying adjacent to the dry summer subtropics have the lowest summer rainfall averages. The absence of high summer temperatures is accompanied by reduced potential evapotranspiration so that natural vegetation does not show a marked influence of seasonal drought. Comparison of the water balance graphs for Birmingham, England, and Athens, Greece, shows the greater summer water need at the latter station (Figure 8.7).

At places in the lee of mountain barriers annual precipitation may be less than 50 cm. Extreme southern Chile, the Otago Basin in southern New Zealand, and the lowlands about the Strait of Juan de Fuca in Washington and British Columbia are examples. The dry summer regime along the North American coast extends farther poleward than in any comparable area in the world. Under the Koeppen classification much of the Puget Sound–Willamette Lowland and southwestern British Columbia would be classified as dry summer subtropical in view of the low summer rainfall and a winter month maximum more than three times that of the driest summer month. Some agricultural land in this area is drained by tiles and ditches in winter and irrigated in summer. Cooler temperatures, a shorter dry season, and a location within the marine regional pattern possibly justify inclusion of these low rainfall areas as anomalous subregions of the marine type.

Some stations of the climatic type have a precipitation maximum (or a secondary maximum) in autumn. The graphs for Paris and Punta

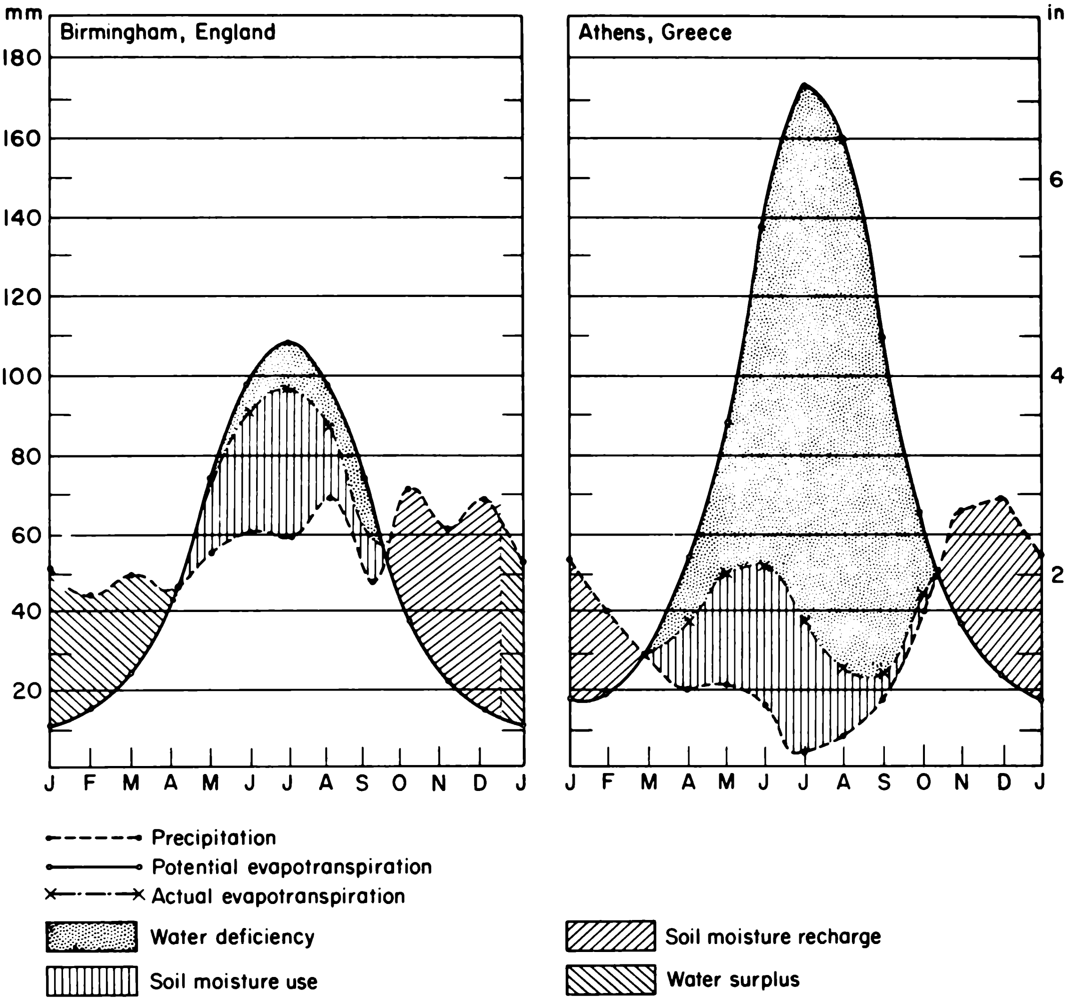


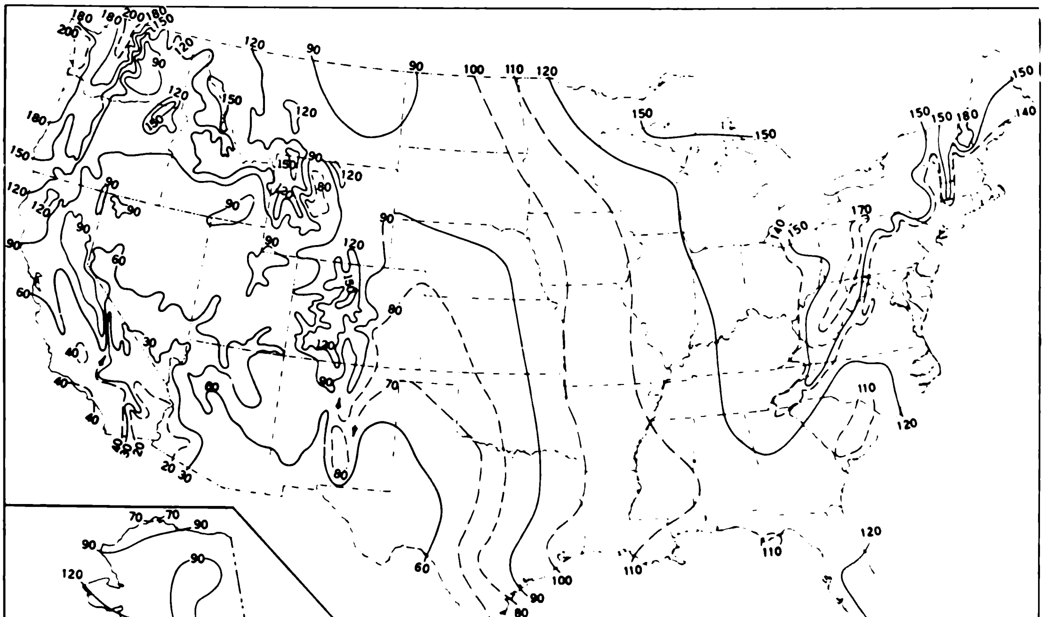
Figure 8.7
Water budget graphs for Birmingham and Athens. (After method of Thornthwaite, based on data computed by C. W. Thornthwaite Associates, Laboratory of Climatology)

Arenas illustrate this regime. October is the rainiest month at Juneau and Sitka, Alaska. Autumn maximums are attributable to the warmer source regions of maritime air masses at that season which give the air a higher moisture content than in late winter or spring.

Two features of the marine climate are the reliability of precipitation and the large number of rain-days, especially in winter. Great departures from the annual or monthly means are uncommon. Many stations have more than 150 rain-days a year. (See Figure 8.8.) Bahia Felix, Chile, near the Straits of Magellan, has an average of 325 rain-days per year; in 1916, rain fell there on 348 days! In such a climate gray, overcast skies with light rain or drizzle are the rule. There may be a number of consecutive days with precipitation, but the total rainfall during such periods is small. Dull, drizzly weather is the price paid by inhabitants of the marine climatic regions for mild winter temperatures. The parade of cyclonic storms brings maritime air with relatively warm temperatures, but it also brings moisture, and often strong winds. Stationary and occluded fronts are common, the occlusion process often taking place along coastal mountain ranges. On the other hand, several fronts may pass over a region in rapid succession without a single rainless day. Needless to say, there are few sunny days in winter. In the lowlands, snow may occur on a few days, but ordinarily it remains for only a short time. Frequently it is "rained off." Only when an invasion of cold polar air follows a snowfall is there a snow cover for more than a day or two.

Summer precipitation is also largely of cyclonic origin. It is more showery and falls on fewer days and yet may reach monthly totals equal to those of winter, the exception being those coasts which come under marked influence of the oceanic highs. Thunderstorms are far less numerous than in the subtropical or the continental climates, for the air which moves in from the high cells tends to be more stable. Most lowland areas average fewer than 10 thunderstorms a year. At higher ele-

Figure 8.8
Mean annual number of days with precipitation of 0.25 mm or more in the United States. (NOAA)



vations, in the mountain areas, total precipitation is, of course, greater and thunderstorm activity is more common. Quite often, conditionally unstable maritime air is cooled sufficiently by orographic lifting to set off convective storms along mountain ranges. In company with the showery precipitation and smaller number of rain-days in summer, there is more sunshine. One of the outstanding seasonal differences in the marine climate is the greater percentage of possible sunshine in summer. In the United States Pacific Northwest, summer is popularly defined as being synonymous with the sunny period. Though of short and variable duration, these "summers" have weather as pleasant as any on earth.

Fog frequency is high in the marine climate. The maximum incidence of fog is in autumn and winter, when warm, stable maritime air drifts across land which has undergone considerable cooling. The marine climate is remarkably free from violent storms such as hurricanes or tornadoes. However, winter winds of gale force are frequent, and, along the coasts, heavy seas are a hazard to shipping and shore installations. The name "roaring forties," used to indicate the fury of the prevailing westerlies in the Southern Hemisphere, is nearly as apt in the Northern Hemisphere. The North Atlantic, Irish Sea, Straits of Magellan, Australian Bight, and Tasman Sea are all infamous in the annals of merchant shipping.

MID-LATITUDE ARID AND SEMIARID CLIMATES

The dry climates of mid-latitudes differ from the tropical arid and semiarid climates in two important respects: average temperatures are lower, and subsiding air masses are not the chief controlling factors. A major factor influencing the arid and semiarid climates of middle latitudes is their location in the continents, far-removed from the windward coasts. Mountain barriers on the windward accentuate the aridity of several areas. The arid and semiarid phases of the dry mid-latitudes differ primarily in the degree of aridity, and they can conveniently be treated together here. This is not to say, however, that the differences in amount or effectiveness of precipitation are unimportant in their effects on crops and vegetation. Three continental regions comprise the principal distribution of these climates: the intermontane basins and Great Plains of western United States and Canada, southern U.S.S.R. and northern China, and western and southern Argentina. In the Northern Hemisphere, the mid-latitude dry climates merge into their tropical counterparts. Temperature is the differentiating criterion and makes itself felt in the demands of evapotranspiration. The transitions between tropical and mid-latitude dry climates are gradual, and boundaries drawn on the basis of temperature values alone are necessarily arbitrary.

The lower annual temperature averages are a response to the effects of higher latitude, and in some cases of altitude as well. Appreciable differences occur within the mid-latitude dry climates as a result of the latitudinal spread. The characteristic temperature curve shows a large annual range, a reflection of continental location. In summer the temperatures are high as continental tropical air masses develop over the heated land surfaces. Because there is a deficiency of moisture little heat is used in evaporation, most of it being used to further warm the subsiding air mass. In winter the prevalent air masses are again continental, but they are associated primarily with outbreaks of polar air. Denver is representative of the mid-latitude semiarid climate; Lovelock, Balkash, and Santa Cruz typify the arid phase (see the accompanying table).

Denver, Colorado 40°N; 1,615 m

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	0	1	4	9	14	20	24	23	18	12	5	2	11
P (mm)	12	16	27	47	61	32	31	28	23	24	16	10	327

Lovelock, Nevada 40°N; 1,212 m

T (°C)	0	3	6	10	15	19	24	23	18	12	5	1	11
P (mm)	16	16	13	11	16	18	5	6	7	11	14	15	146

Balkash, U.S.S.R. 47°N; 423 m

T (°C)	-15	-13	-5	8	16	22	24	22	15	6	-5	-12	5
P (mm)	10	8	10	11	9	14	11	9	4	8	9	12	115

Santa Cruz, Argentina 50°S; 12 m

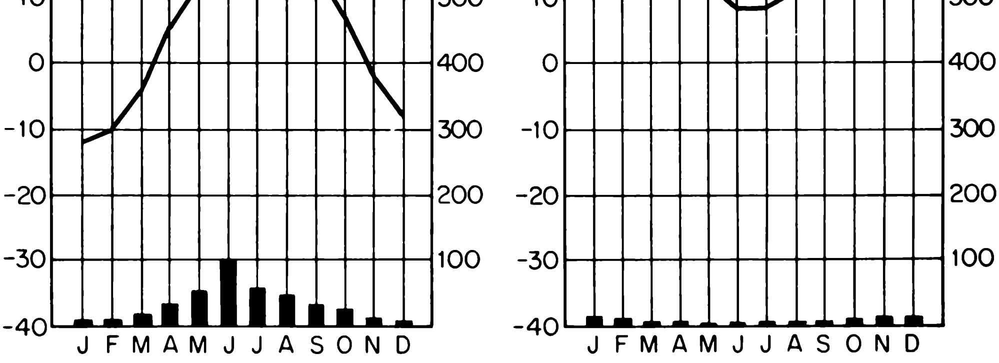
T (°C)	14	14	12	9	5	2	2	4	6	10	12	14	8
P (mm)	21	16	20	17	25	18	16	16	12	7	15	18	201

Note the contrasting temperatures at these representative stations. Deep in the Asian continent, Balkash has an annual range of 39C°, whereas Santa Cruz, on the southern Argentina coast, has a range of only 12C°. Patagonia lies to the leeward of the Andes and has a marine temperature regime, but it receives little moisture from the prevailing westerly flow of air. Only occasionally does cyclonic activity bring easterly winds, and the continent is too narrow to induce a thermal low-pressure center with monsoonal circulation onto the land in summer. The modified marine influence of southern Argentina is unique in the dry mid-latitude climates.

Extreme temperatures range from well below −40°C in winter to above 40°C in summer in these climates. The lowest minimums are recorded in the northern Great Plains and in Siberia. The highest maximums occur at the lower-latitude margins of the climates. With respect to temperature, a summer day in certain parts of the dry mid-latitudes may be uncomfortably like the tropical deserts. Diurnal ranges are great, however, and especially at the higher elevations the nights can be pleasantly cool. Daytime heating and nocturnal cooling effect a large diurnal range, just as in the tropical arid and semiarid climates. Although the same is true to some extent in winter, the mid-latitudes have much lower monthly averages, and invasions of polar continental air bring severe freezing temperatures not found in the tropics.

In common with the dry climates of the low latitudes the mid-latitude arid and semiarid climates have deficient precipitation. The minimum amount of precipitation that will support a vegetation association of a given density is, however, lower than in the tropics, for the temperatures and the consequent evaporation and transpiration are lower. Annual rainfall of 15 to 20 cm may support steppe vegetation in the cool mid-latitudes but would be adequate only for desert scrub in the tropics. It is ultimately on the basis of the ratio of precipitation to potential evapotranspiration that the arid and semiarid phases of the dry mid-latitude climates are best differentiated. Desert vegetation is characteristic of the arid climate and steppe grassland predominates in the semiarid regions. Accordingly, the climates are also well known by their associated vegetation as mid-latitude desert and steppe, respectively.

Figure 8.9
Climatic graphs for Dickinson, mid-latitude semiarid, and San Juan, mid-latitude arid.



Dickinson, North Dakota	47°N; 750m	San Juan, Argentina	32°S; 630m
Mean annual temperature	5°C	Mean annual temperature	17°C
Annual temperature range	33°	Annual temperature range	18°
Mean annual precipitation	392 mm	Mean annual precipitation	86mm

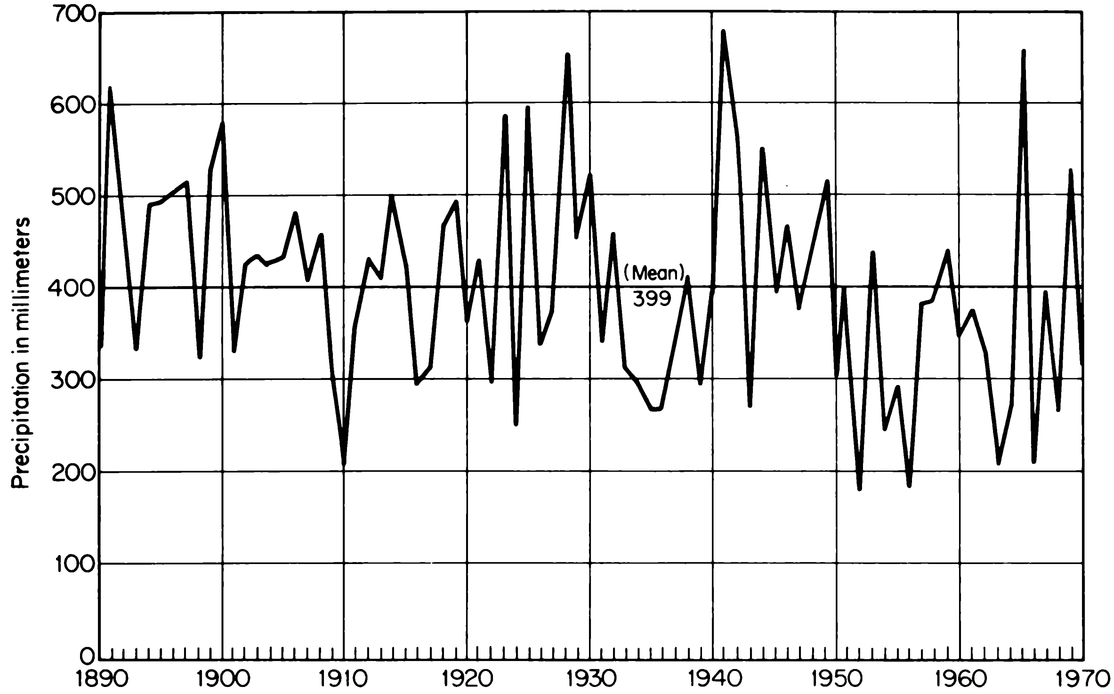


Figure 8.10
Variations in annual precipitation at a station in the mid-latitude semiarid climate near Springfield, Colorado, 1890–1970.

As in the dry tropical climates, in the mid-latitude counterparts there is no general rule governing the regime of precipitation. The margins near the dry summer subtropics have winter maximums; toward the humid continental climates summer maximums prevail, and where marine influences are marked, as on the east coast of Argentina, there is no great variation from month to month. The graph for San Juan illustrates the winter maximum in the arid type; Dickinson, North Dakota, has an early summer maximum (Figure 8.9). Absence of a distinct seasonal maximum is shown in the data for Santa Cruz and Balkash.

A characteristic of the precipitation record in all dry climates is the great variability from year to year. (See Figure 8.10.) In general, departures from the long-term mean increase as the mean decreases.

In a study which has become a classic in geographic literature, Henry M. Kendall pointed out that, on the Great Plains in the period 1915–24, humid years were experienced as far west as the Rocky Mountains and dry years occurred as far east as Minnesota. The significance of precipitation variability in the semiarid climate is far greater than in the arid. A 100 percent increase of rainfall in a given year in an arid climate with an average of 7 cm is not likely to revolutionize vegetation and human settlement. Agriculture is geared to irrigation, if practiced at all. In the steppe regions, however, a series of wetter-than-average years gives a false impression of agricultural potentialities, and settlers are tempted to institute land uses not actually suited to the long-term climatic pattern. In no climate is there more danger in the use of statistical averages than in the mid-latitude semiarid type. The dry years are as important as the "normal" years, if not more so, in determining the vegetation association and the optimum patterns of land use.

The meager winter precipitation in the dry mid-latitudes is frontal and comes with occasional cyclones, which replace the continental high-pressure centers. It may come as snow, and high winds accompanying polar outbursts in the wake of a cold front produce blinding snowstorms, known as *blizzards* on the Great Plains. The infrequent cyclonic disturbances are sandwiched between periods of cold and generally clear weather. The chinook, or foehn wind, is experienced over the northern Great Plains in winter and spring. A similar wind which descends from the Andes into western Patagonia is known as the *zonda*. In the Soviet Union the *sukhovey* is a warm, dry wind associated with summer anticyclones over southern Siberia. Surface wind speeds are generally high in these climates because of the large expanse of plains and plateaus. Immediately at the surface, the absence of trees is also a factor permitting free flow of air.

Summer weather is influenced by thermally induced low-pressure centers over the continents. Although this may tend to produce monsoon-type winds blowing onto the continent from water bodies, mountain barriers and the distance of the dry climatic regions from the sea limit the amount of moisture that reaches the interior. Summer rainfall comes with scattered thundershowers. All too often a dry period is broken by a violent cloudburst-hailstorm which does as much damage to crops as drought. In the United States, tornadoes sometimes form along belts of frontal thunderstorms, especially in the warmer half of the year.

HUMID CONTINENTAL WARM SUMMER CLIMATE

Extensive areas with humid continental climates occur only in the Northern Hemisphere; the Southern Hemisphere has mostly a water surface at these latitudes. As indicated by their names, the two types of humid continental climate differ primarily with respect to the length and intensity of the summer season. They are considered separately here to facilitate emphasis on their differing temperature characteristics.

The humid continental warm summer climate lies in three areas between latitudes approximately 35° to 45° north. In the United States it is roughly coextensive with the Corn Belt, bordering the semiarid climate of the Great Plains on the west and running eastward to the Atlantic with an interruption at the Appalachians. In southeastern Europe it is centered on the Danube Basin and includes portions of Hungary, Romania, Yugoslavia, and Bulgaria. The long belt of dry climate in central Asia separates the second area from the third, which is found in Manchuria, northeastern China, Korea, and northern Honshu. In winter, polar continental air masses dominate, bringing much colder weather with intervening surges of tropical maritime air in the warm sectors of cyclones; in summer, maritime and continental tropical air masses bring high temperatures and rainfall maximums. This is truly the battleground of polar and tropical air masses. There are cyclonic "skirmishes" within the larger pattern of seasonal "warfare."

Annual temperature averages mask the large annual ranges that are characteristic of the humid continental warm summer climate. Summer is hot; winter is cold. Mean monthly temperatures for June, July, and August are near or above 20°C. The average length of the frost-free season ranges from about 200 days in the south to 150 days along the northern margins. Yet January averages are typically below 0°C. St. Louis, near the boundary of the humid subtropics, has a January average of 0°C but a July average of 26°C. Harbin, near the boundary of the cool summer phase in Manchuria, has −20°C as a January average and 23°C in July (see accompanying table).

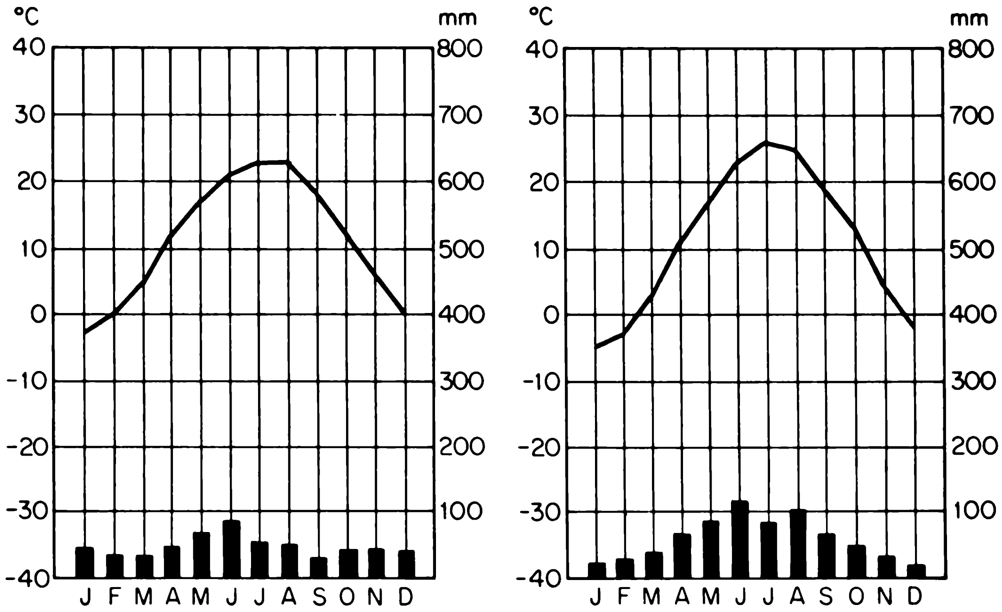
St. Louis, Missouri 39°N; 172 m

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	0	2	6	13	19	24	26	25	21	15	7	2	13
P (mm)	50	52	78	94	95	109	84	77	70	73	65	50	897

Harbin, Manchuria 46°N; 143 m

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	−20	−16	−6	6	14	20	23	22	14	6	−7	−17	3
P (mm)	4	6	17	23	44	92	167	119	52	36	12	5	577

A characteristic of continental climates is a great annual range of temperature. This is expressed in the warm summer type not only by high monthly means in summer but also by extreme maximums above 35°C. Furthermore, diurnal ranges are small in summer so that the nights are often uncomfortably warm. Conditions of high humidity and high temperature are similar to those in the humid subtropics. A day and night in Des Moines, Iowa, may be much like the same period in New Orleans, Louisiana, or Montgomery, Alabama, for temperatures do not decrease rapidly with latitude in this climate in summer. Winter is quite another matter. Not only are the monthly averages and the extreme minimums much lower, but the decrease in these values with increasing



Bucharest, Romania	44°N; 82 m
Mean annual temperature	11°C
Annual temperature range	26°
Mean annual precipitation	578 mm

Omaha, Nebraska	41°N; 304 m
Mean annual temperature	11°C
Annual temperature range	31°
Mean annual precipitation	700 mm

Figure 8.11
Climatic graphs for Bucharest and Omaha, humid continental warm summer climates.

latitude is appreciable. The cold continental polar outbursts which bring the very low temperatures to the northern parts of the climatic region may not reach the southern margins, or, if they do, they are modified somewhat as they move to lower latitudes. Thus, typical annual minimum temperatures in northern Iowa are on the order of −30°C, whereas in southern Missouri the mean annual minimum is −20°C. East of the Great Plains in North America temperatures increase southward at a faster rate in winter than in summer. The influence of the large land mass in Asia results in even lower winter temperatures in the humid continental warm summer climate of Manchuria.

Annual precipitation in this climate varies from 50 to 125 cm. The amount typically decreases toward the northern latitudes and toward continental interiors. Spring or summer is the season of maximum rainfall. There are occasional frontal passages, but convective showers account for most of the warm-season rain. Scattered thunderstorms and a high percentage of possible sunshine produce summer weather similar to the humid subtropics. A maximum of rainfall in the summer half of the year is most marked in the interior areas and in eastern Asia, where the monsoonal influence is an important factor. For example, in July Harbin has 16.7 cm but in January only 4 mm. A characteristic of the precipitation regimes of interior locations in this climate is the spring or early summer maximum, usually in May or June. (See Figure 8.11.) In spring the air masses are still cold from the north and the contrasting tropical maritime air comes from water that has reached its lowest temperature for the year. The land heats rapidly, however, and the consequent warming of lower layers of advancing maritime air increases instability and leads to numerous convective showers. Although thunderstorms continue throughout the summer the differences in temperature between the land and the upper air become less and the degree of instability is therefore lessened. Sometimes hail accompanies thunderstorms in this climate. In the Mississippi Valley tornadoes are an additional hazard, primarily in the warm months.

Figure 8.12
Agricultural landscape in the humid continental warm summer climate. The humid continental climatic regions are noted for grain, hay, and livestock production in both North America and Eurasia. This scene is in eastern South Dakota. (Bureau of Reclamation, U.S. Department of the Interior)

Except on the east coast of the United States, the winter months usually have much less precipitation, a feature which is particularly evident in eastern Manchuria and northern China, where a monsoonal regime prevails. A part of the winter precipitation comes as snow, the proportion increasing with latitude and altitude. It is primarily of frontal origin, and the blizzards behind the cold fronts may extend into the humid continental warm summer climate from the west or north. Average annual snowfall varies from 25 to 100 cm. Frontal weather in winter may also bring rain, occasional fog, and sometimes ice pellets or freezing rain. Air mass thunderstorms can develop in the warm sectors of cyclones whose paths follow the southern margins of the climatic type. In general, winter is much less sunny than summer, a fact taken for granted by residents in the areas with humid continental climate. Nevertheless, conditions are in marked contrast with the wet-and-dry tropics, which also have a precipitation maximum in summer but which have more sunny weather in winter.

Probably the best-known feature of winter weather in the humid continental climates is its changeability. Temperatures may drop from above 15°C in the warm sector of a cyclone to −20°C behind the cold front in a period of a day or two, and, although extremes of such magnitude are not the rule in winter cyclones, sudden changes typify most storms. The passage of cyclones becomes more frequent in autumn, but there may be warm, sunny weather between the first two or three cyclones. In the United States the weather associated with stable air in October and November is often called "Indian summer," and since it often follows frosts, the fall colors of deciduous trees add a great deal to the pleasantness of these periods. Just as summer retreats with rearguard delaying actions, so in spring there are late frosts and snowstorms which punctuate the erratic poleward retreat of cyclonic storms. Custom decrees that these late spring storms be termed "unseasonal"; they are in fact a characteristic of the climate.

HUMID CONTINENTAL COOL SUMMER CLIMATE

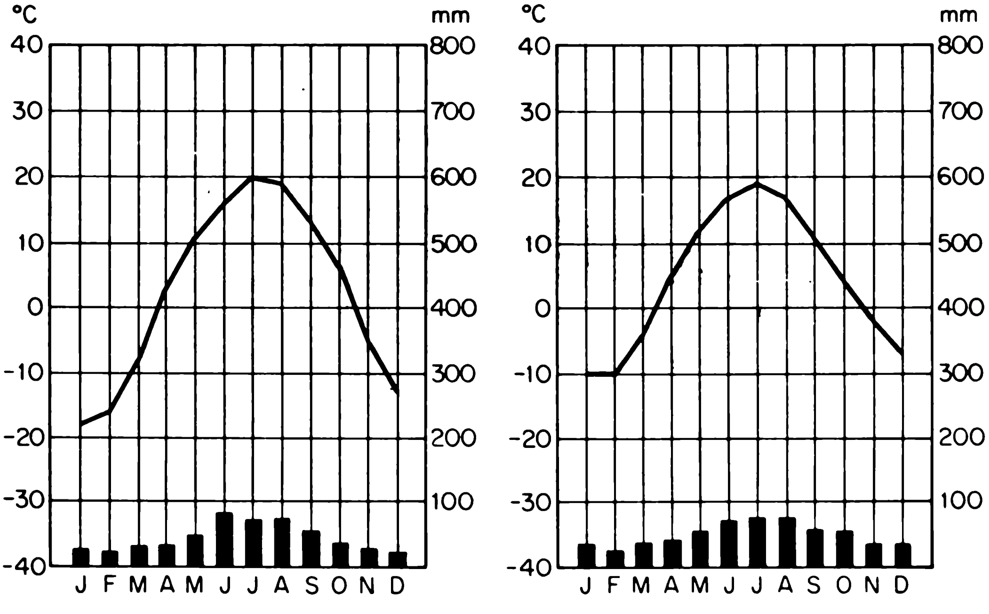
The humid continental cool summer climate is like the warm summer type in so many ways that it can be discussed by comparing the two. The most obvious distinguishing features have already been mentioned, the cooler summer and shorter growing season, which result from location at higher latitudes. The climatic region lies north of the warm summer type in North America, eastern Europe, and the far east of Asia. In each case it has wider longitudinal extent than the humid continental warm summer type. In North America it roughly straddles the

Canadian–United States border, with a poleward extension into central Alberta on the west and an eastern limit at the Atlantic Coast in the Maritime Provinces and New England. In Europe and the U.S.S.R. it takes the shape of a long isosceles triangle with southern Scandinavia, Poland, and Czechoslovakia at the base and the vertex deep in central Siberia. In Asia it extends from northern Manchuria out to Hokkaido and southern Sakhalin.

Being in higher latitudes, the cool summer type of humid continental climate is affected by polar air masses for a longer season. Incursions of tropical continental or tropical maritime air are more frequent in summer. On the east coasts, maritime air is drawn onto the continents as cyclones approach the sea.

Temperatures are generally lower both in winter and summer than in the warm summer type. Summer extremes may exceed 35°C in a "heat wave" of tropical continental air, but these are offset by cool periods, often with cloudiness. Winnipeg and Moscow have July means of 20°C and 19°C, respectively, in contrast with July means of 26°C at Omaha and 23°C at Bucharest (Figure 8.13).

The frost-free season is generally less than 150 days in length,



Winnipeg, Manitoba	50°N; 240m
Mean annual temperature	2°C
Annual temperature range	38°
Mean annual precipitation	517mm

Moscow, U.S.S.R.	56°N; 156m
Mean annual temperature	4°C
Annual temperature range	39°
Mean annual precipitation	575mm

Figure 8.13

Climatic graphs for Winnipeg and Moscow, humid continental cool summer climates.

except for coastal locations, and it decreases in the higher latitudes. It is significant, however, that with the increase of latitude the summer daylight periods are longer; this tends to offset the short season to some extent insofar as crop production is concerned. Thus Edmonton, Alberta (lat. 53°45′ N), has more than 17 hours, compared with less than 15 hours at St. Louis, Missouri (lat. 38°35′ N), in the warm summer climatic type.

In the winter the days are increasingly shorter as one proceeds northward and the contrast between the seasons is correspondingly accentuated. Annual ranges of temperatures are even greater than in the humid continental warm summer climate. Winnipeg has a range of 38C°; at Moscow the range is 29C°. These large annual ranges result from a combination of latitude and interior continental location. Winters are severe; typically the January average is well below 0°C, and for some interior stations it is below −15°C. Note, however, that the climate is identified primarily by its summer rather than its winter temperature characteristics. Harbin, Manchuria, for example, has a January mean temperature of −20°C but a warmer and longer summer than in the humid continental cool summer type of climate. Extreme minimums in the cool summer type fall to −45° or −50°C but they are not an annual occurrence. A series of several days having temperatures below −15°C can be expected in most winters as polar or arctic air lies over the continents. Short-term changes of temperature are greatest in winter, and they are more marked in connection with the nonperiodic cyclonic disturbances than with the diurnal cycle of insolation.

Annual precipitation in the humid continental cool summer climate is usually less than in the warm summer type. The areas within the climatic region are farther from the sources of maritime tropical air and they are influenced for a greater part of the year by drier polar air masses. Annual amounts for most stations range from 35 to 70 cm, but the temperatures are lower and a given amount of summer precipitation is more effective than in the warm summer type. Regimes of precipitation are similar to those of the warm summer type, the maximum coming in summer in the interiors. Toward the east coasts of North America and Asia and in Japan, the summer maximums become less pronounced and autumn maximums are more characteristic. Summer thunderstorms are not so common, but summer frontal passages occur more frequently. In the winter season the proportion of snow is greater and snow lies for longer periods. Prolonged snow cover is a factor which inhibits the warming of southward-moving air masses, and, consequently, contributes to the severity of winter conditions.

The proximity of oceans along the east coasts of North America and Asia considerably modifies the humid continental climate. Southerly

Eastport, Maine 46°N; 24 m

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	−5	−5	−1	5	9	13	16	17	14	9	4	−3	6
P (mm)	104	87	93	89	78	90	78	73	90	92	114	95	1,084

Nemuro, Japan 43°N; 26 m

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	−5	−6	−2	3	7	10	14	18	16	11	5	−1	6
P (mm)	49	40	77	77	99	97	104	106	152	124	92	63	1,081

and easterly winds in the cyclonic circulation bring maritime air to these regions, resulting in increased precipitation and moderated temperatures, especially in winter. Eastport, Maine, has 108 cm of precipitation fairly well distributed through the year. Nemuro, on Hokkaido in northern Japan, has the same annual average but a September maximum (see the accompanying table). The peculiar features of the humid continental cool summer climate on the east coasts have led some authors to classify it as a distinctive "modified humid continental" or "New England-Hokkaido" type. On the coasts of New England and the Canadian Maritime Provinces the precipitation regime varies less than in the interior; Cape Race, Newfoundland, has a November maximum. Mean summer temperatures and annual ranges are both appreciably lower than inland. The Maine coast has an exceedingly steep gradient of annual temperature ranges away from the coast. At Eastport the range is 22C°, whereas at Woodland, only 50 km inland, the range is 30C°. The ocean not only modifies temperature ranges, but also produces a longer

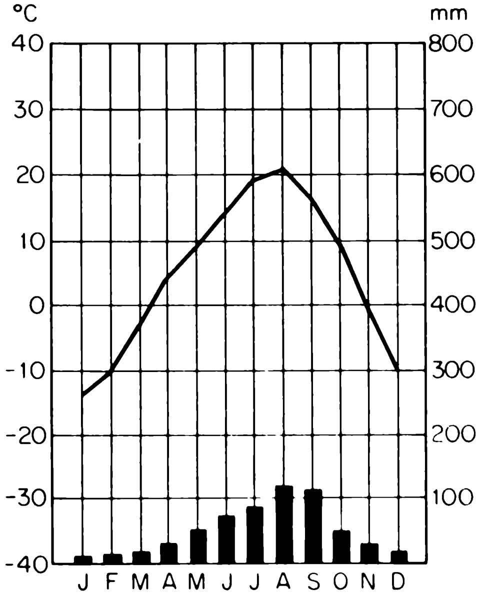


Figure 8.14
Climatic graph for Vladivostok, humid continental cool summer climate.

Vladivostok, U.S.S.R.	43°N; 29m
Mean annual temperature	4°C
Annual temperature range	35°
Mean annual precipitation	598mm

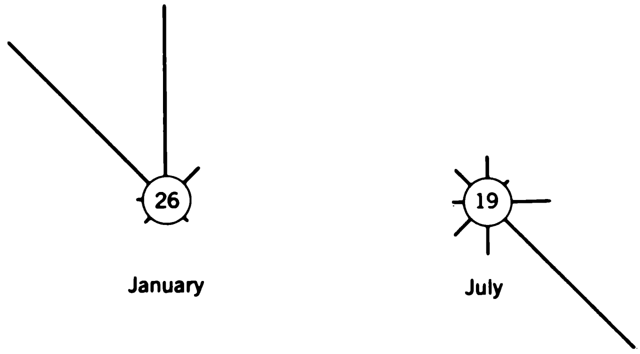


Figure 8.15
Wind roses for Vladivostok, U.S.S.R.

frost-free season. A further characteristic of the coast of New England and the Maritimes is the frequency of gales associated with the succession of cyclones. On rare occasions tropical hurricanes reach this area. Off the coast of Canada the tropical water of the Gulf Stream meets the cold Labrador Current and fogs are frequent in the overlying air. Light winds carry the fog onto the coasts, but it rarely penetrates far inland. The greatest fog frequency is in summer months and over the Grand Banks east of Newfoundland.

Maritime effects in northern Japan are basically similar to those in New England, but the persistent continental air masses act as a barrier to oceanic air in winter, resulting in an extension of the continental characteristics to the east coast of mainland Asia. The mountainous islands receive moisture in winter from the Sea of Japan and thus do not exhibit the usual dry winter regime of the mainland. Note that Nemuro has a September maximum of rainfall which coincides with the typhoon season. Comparison of the data for Nemuro and Vladivostok reveals the effect of an island location in producing a smaller annual temperature range and a more even distribution of precipitation (see Figure 8.14). The winter climate of Vladivostok is essentially continental; in the month of January winds are almost entirely from the north or northwest (Figure 8.15).

Additional Readings

Clarke, John F., "Some Climatological Aspects of Heat Waves in the Contiguous United States," *Environ. Res.*, 5, 1 (1972), 76–84.

Climatic Atlas of the United States. Washington: U.S. Department of Commerce, Environmental Science Services Administration, 1968.

Dury, G. H., "High Temperature Extremes in Australia," *Ann. Assoc. Am. Geogrs.*, 62, 3 (1972), 388–400.

FRISBY, E. M., " A Study of Hailstorms of the Upper Great Plains of the North American Continent," *Weatherwise,* 17, 2 (1964), 68–74.

GENTILLI, J., ed., *Climates of Australia and New Zealand,* Vol. 13, World Survey of Climatology. New York: American Elsevier Publishing Company, 1971.

KENDALL, HENRY M., "Notes on Climatic Boundaries in the Eastern United States," *Geog. Rev.,* 25, 1 (1935), 117–24.

KENDREW, W. G., AND B. W. CURRIE, *The Climate of Central Canada.* Ottawa: Queen's Printer, 1955.

KIMBLE, GEORGE H. T., *Our American Weather.* New York: McGraw-Hill Book Company, 1955.

LICHTENSTEIN, E. R., AND M. L. SCHWARZKOPF, "Squall-lines in Argentina," *Weather,* 21, 5 (1966), 181–86.

RODDA, J. C., "Rainfall Excesses in the United Kingdom," *Trans. Inst. Brit. Geogrs.,* 49 (1970), 49–60.

chapter nine

Climates Dominated by Polar and Arctic Air Masses; Highland Climates

The most widespread variations in temperature on the earth are those resulting from latitudinal and altitudinal variations in the net gain of heat. An increase in either latitude or altitude is normally accompanied by decreasing temperatures, and distinctive features of the climates of high latitudes or high altitudes arise primarily from the low temperatures. Temperature and precipitation characteristics differ between the two climatic groups, however. We shall consider first the cold climates of high latitudes.

TAIGA CLIMATE

The term *taiga* is of Russian origin and refers to the northern continental forests of North America and Eurasia. It has been used also to designate the distinctive climate that prevails in these forest regions. Names such as "subarctic" or "boreal forest" have been applied to the taiga climate.

Two large world regions embrace the taiga climate. In North America it extends in a broad belt from western Alaska to Newfoundland; in Eurasia from Norway to the Kamchatka Peninsula. In both regions the climatic type reaches farther south on the eastern margins of the continent than on the west, reflecting the effect of the large land masses. These regions are the sources of polar air masses, and for most of the year their weather is dominated by cold, dry, stable air. As a result annual temperature ranges are large and precipitation is meager.

The mean July temperature at Yakutsk, in east central Siberia, is 20°C, but the January average is −43°C, making the annual range 63C° (see the accompanying table). Ranges of 30C° or more are typical in the interior locations; conditions are moderated near the oceans, particularly on west coasts. Tromso, Norway, has a mean annual range of only 15C°.

Yakutsk, U.S.S.R. 62°N; 103 m

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	-43	-37	-23	-7	7	16	20	16	6	-8	-28	-40	-10
P (mm)	7	6	5	7	16	31	43	38	22	16	13	9	213

Tromso, Norway 70°N; 24 m

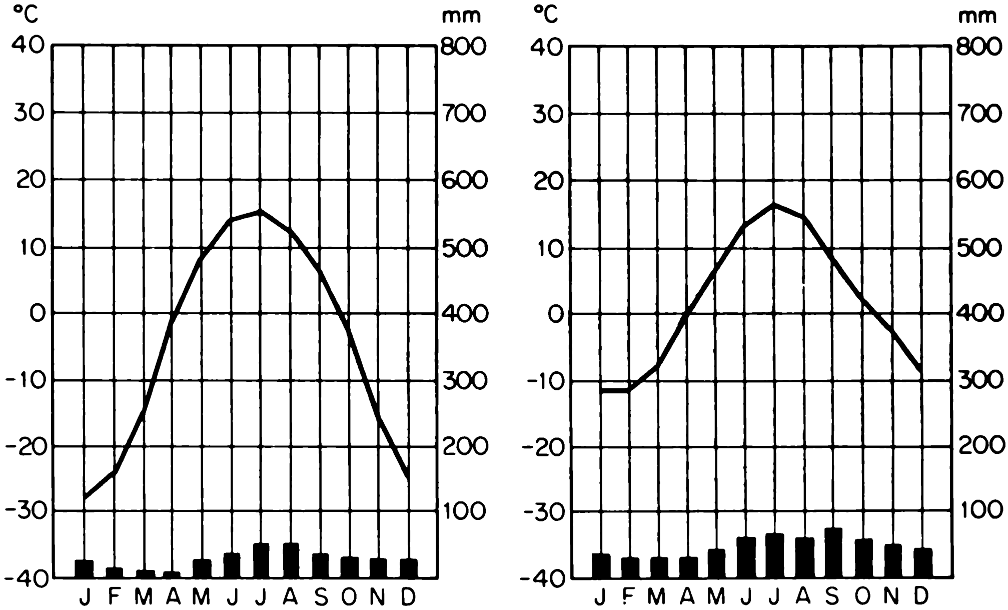
T (°C)	-3	-3	-2	1	5	9	12	11	8	4	0	-1	3
P (mm)	118	94	113	75	65	57	56	83	115	131	97	115	1,119

Fairbanks, Alaska 65°N; 134 m

T (°C)	-24	-19	-13	-1	8	15	15	12	6	-3	-16	-22	-3
P (mm)	23	13	10	6	18	35	47	56	28	22	15	14	287

In the short summer the days are long, and maximum temperatures frequently rise above 25°C. Yakutsk has recorded a temperature of 39°C. Fairbanks has an extreme maximum of 34°C. At such stations a small diurnal range might be expected as a result of the long daylight periods, but even where there are 24 hours of daylight an appreciable difference exists between the amount of effective insolation at noon and that at midnight, when the angle of the sun is low. Thus, diurnal ranges are commonly on the order of 10° or 15C°. Marine influences moderate the diurnal range considerably. The frost-free season in the taiga varies with the latitude and with distance from the sea. Inland it lasts for 50 to 90 days, but there is always the risk of summer frost. The short duration of the frost-free season is balanced to some extent by the longer days, so that certain hardy crops can mature in a shorter calendar period in the taiga than in lower latitudes. Not only is the summer short, but it also ends abruptly. Autumn is often said to begin with a hard freeze, last a few days, and end with the first snow of winter.

Winter is the dominant season. Except on favored coasts, six to eight months have mean temperatures below 0°C, and many stations have three or four months with means below −15°C. (See Figure 9.1.) Furthermore, average daily *maximum* temperatures in winter are gener-



Dawson, Yukon	64°N; 323m
Mean annual temperature	-5°C
Annual temperature range	43°
Mean annual precipitation	323mm

Archangel, U.S.S.R.	65°N; 13m
Mean annual temperature	1°C
Annual temperature range	28°
Mean annual precipitation	539mm

Figure 9.1
Climatic graphs for Dawson and Archangel, taiga climates.

ally well below freezing. Under the dominance of the continental anticyclones the weather is usually clear, but the daylight periods are short or entirely absent and outgoing radiation far exceeds insolation. The lowest official surface temperature on record in the climatic type is −68°C at Verkhoyansk in eastern Siberia. It occurred on the 5th and again on the 7th of February, 1892. At Oimekon, about 650 km to the southeast, the temperature record is shorter, but in recent years it has been colder than Verkhoyansk. An unofficial −78°C was reported at Oimekon in 1938, but details of the date and exposure of the thermometer are unknown. In North America the lowest official temperature is −63°C, recorded at Snag, Yukon Territory, on February 3, 1947. (See Figure 9.2.) Although these are exceptional temperatures, they indicate the extreme cold which frequently envelopes the taiga in winter. Under the calm conditions that sometimes accompany the extremely low temperatures the dry air does not produce the same physiological chilling as moister air would at appreciably higher temperatures. With wind, however, chilling effects rapidly increase and powdered snow may pervade the air, reducing visibility to zero in a "whiteout."

Except along the seaward margins, precipitation in the taiga cli-

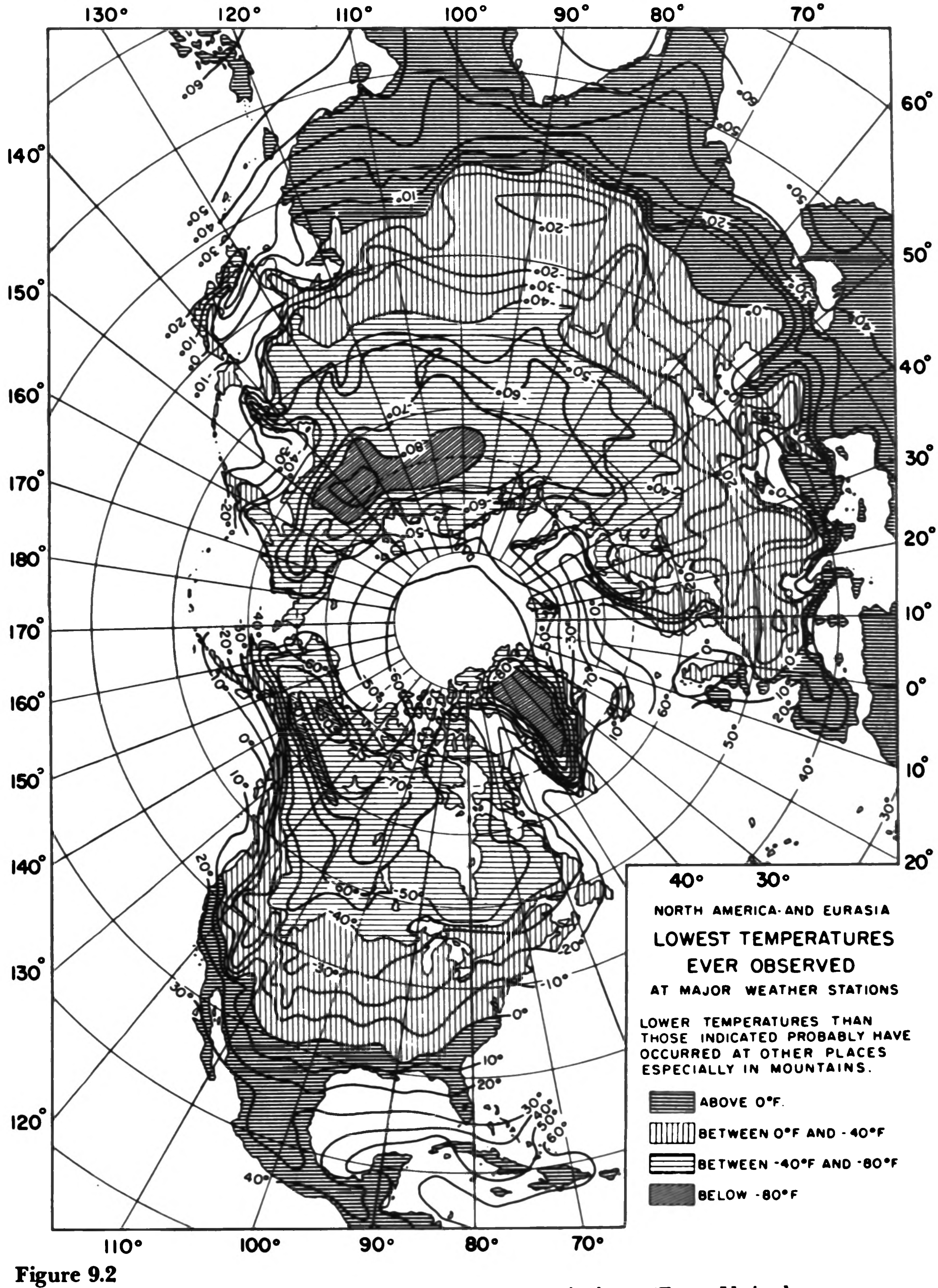


Figure 9.2
Extreme minimum temperatures in the Northern Hemisphere. (From United States Air Force, Geophysics Research Directorate, *Handbook of Geophysics*, Revised Edition, published by The Macmillan Company, 1960)

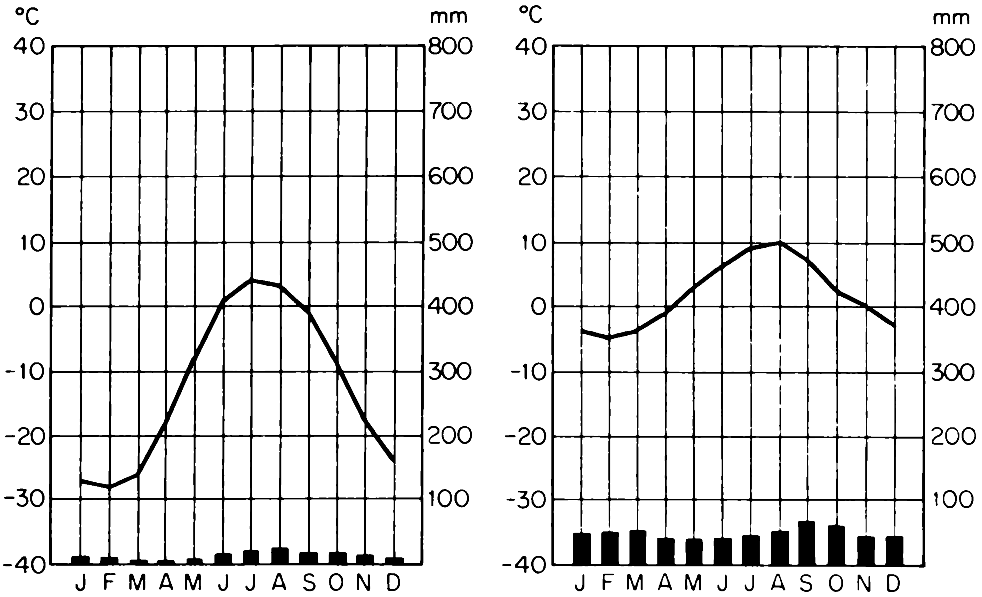
mate is generally less than 50 cm annually. It is mostly of cyclonic origin and has a definite summer maximum, again with the coasts excepted. (Compare the data for Tromso and Fairbanks.) At interior locations rainfall is often inadequate to meet the demands of potential evapotranspiration during the long daylight periods of summer, so that dust forms and irrigation may even be necessary in the few areas where crops are raised. Winter snow actually accounts for the smaller part of annual precipitation in most of the taiga, although it may accumulate for several months.

Fog and low stratus clouds are fairly common along the coasts of this climatic region, being especially prevalent in the Aleutian Islands. Ice fogs may form over snow surfaces in winter when temperatures are extremely low.

TUNDRA CLIMATE

Like taiga, *tundra* is a vegetation term which has also been applied to the associated climate. The tundra climatic region lies north of the taiga and extends along the arctic coasts of North America and Eurasia. For the most part it is found north of the Arctic Circle, but in eastern Canada and in eastern Siberia it extends into lower latitudes in keeping with the continental influence. Northern Iceland, coastal Greenland, and many islands in the Arctic Ocean also have this type of climate. The climate does not occur in the Southern Hemisphere except on a few islands in the Antarctic Ocean and small areas on Antarctica, because of the predominance of water surface in the corresponding latitudes.

For most of the year the tundra climate is under the influence of continental polar or continental arctic air masses. In summer the air may have maritime characteristics as a result of the melting of sea ice and more frequent advances of air from oceanic sources in lower latitudes. Even in winter, islands and coasts experience limited marine influences. These effects are the basis for a possible subdivision, the polar marine climate. Although the long periods of ice cover on the Arctic Ocean render it less effective as a moderating influence than oceans at lower latitudes, it nevertheless does affect the tundra climate. Even in winter, when entirely frozen over, the Arctic Ocean has enough heat to modify temperatures slightly. For this reason the extreme minimum temperatures occur farther south within the continents, where loss of heat by radiation is great, rather than at the higher latitudes on the coast. The boundary between sea-ice and water is probably one of the most significant factors in the weather of the arctic. Along this boundary the greatest horizontal temperature gradients occur, and it is consequently a zone of frequent storms.



Barrow, Alaska	71°N; 9m
Mean annual temperature	-12°C
Annual temperature range	32°
Mean annual precipitation	110mm

Vardo, Norway	70°N; 15m
Mean annual temperature	2°C
Annual temperature range	15°
Mean annual precipitation	545mm

Figure 9.3
Climatic graphs for Barrow and Vardo, tundra climate.

Mean annual temperatures in the tundra are normally below 0°C, and the annual range is large. Barrow, Alaska, has an annual mean of −12°C, but the July average is 4°C and that for February is −28°C (Figure 9.3). The range of 32C° reflects the marked continental nature of the climate in spite of the station's location on a coast. Vardo, Norway, at a comparable latitude, experiences a greater maritime influence, having an August mean of 10°C, a February mean of −5°C, and an

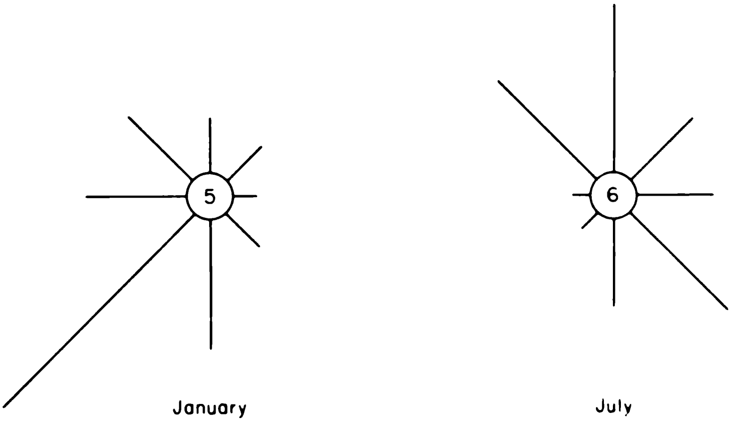


Figure 9.4
Wind roses for Vardo, Norway.

annual range of only 15C°. The more moderate temperatures at Vardo are due in part to the effects of the North Atlantic Drift, which creates positive temperature anomalies in the far northern Atlantic. Prevailing southerly and westerly winds in winter augment the marine influence (see Figure 9.4). There is no comparable movement of warm water through the Aleutian Island chain and Bering Strait into the Arctic Ocean north of Alaska.

Summer temperatures in the tundra are reminiscent of mild winter weather in the mid-latitudes; the average temperature in the warmest month is below 10°C. The extremely long daylight period is not enough to overcome the negative effects of low sun altitude and the ice-choked polar seas. On rare occasions maximum temperatures of over 27°C have been recorded, but a more representative maximum is 15° to 18°C. On the other hand, frosts may occur in any month. Most stations record only two to six months with average temperatures above 0°C. The exceptions are those stations in especially favored coastal locations near the southern margin of the climatic types.

The long, cold winters of the tundra do not differ greatly from those in the taiga. Six to 10 months have mean monthly temperatures below freezing. The high latitude results in a long period when the sun appears above the horizon each day for only a short time or not at all. During this period, the loss of heat by radiation is far in excess of insolation, but the region is not as dark as might be supposed. Cold weather is accompanied by clear skies, and the long nights are illuminated by the moon, stars, twilight, and displays of the aurora borealis. Extreme minimum temperatures are slightly higher than in the taiga because of the adjacent sea, although the difference between, say −45°C and −50°C, is of doubtful importance to human beings in the two regions.

Annual precipitation is less than 35 cm over most of the tundra, and it comes largely with cyclonic storms in the warmer half of the year. In the extreme northern reaches of the Atlantic, however, the warmer sea temperatures and a persistent upper-level trough of converging air streams combine to produce greater amounts of precipitation. (See map of world distribution of precipitation, in Chapter 3.) Thus Ivigtut on the southwest coast of Greenland (see accompanying table) and Angmagssalik on the east coast have 134 cm and 78 cm, respectively. On the west coast Godthaab has only 46 cm.

Ivigtut, Greenland 61°N; 5 m

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	−5	−4	−3	0	5	8	10	9	6	2	−2	−4	2
P (mm)	86	125	86	81	99	100	75	109	172	187	144	77	1,340

Snow contributes a greater proportion of the precipitation in the tundra than in the taiga. Wet summer snows are common along the poleward margins. Winter snowfall ordinarily does not exceed 150 to 200 cm and is probably not more than 65 cm in the northern areas. Snow depths in winter are somewhat greater in northeastern Canada and the lands bordering the North Atlantic than in northern Alaska, northwestern Canada, and Siberia. In the former areas, wintertime cyclonic storms are more common. With the low temperatures, the snow is powdery and drifts easily and is consequently difficult to measure accurately. In the regions of scanty snowfall, large areas may be swept bare while others are covered by deep drifts. Blowing snow frequently reduces visibility to a few meters.

Maximum storminess in the tundra occurs in autumn and in spring. In autumn there are appreciable differences between temperatures of the land and the unfrozen sea, and in spring the zone of cyclonic storms is best developed over the region. Some maritime stations have late summer or autumn precipitation maximums that are related to this seasonal distribution of storms. The precipitation graph for Vardo (Figure 9.3) illustrates this effect.

No general statement can be made about cloudiness and fog in the tundra; the amount varies with storm activity and location with respect to water bodies. Fog and low stratus are prevalent along the seacoasts in the warmer months. Where cold water and drift ice move southward underneath warmer humid air, as along the Labrador Coast and in the Bering Strait, summer fog reaches its maximum. Winter skies are comparatively clear, although ice fog sometimes occurs.

POLAR CLIMATE

In the polar climate mean monthly temperatures are all below 0°C and vegetation is entirely lacking. Snow and ice or barren rock covers such areas. The polar climate and the associated icecaps predominate over most of Greenland, the permanent ice of the Arctic Ocean, and Antarctica. Less is known about this climate than any other because of limited exploration in these areas and the dearth of climatic data. The widely held concept of a permanent anticyclone or high at each of the poles has not been borne out by observations, for the high appears to be weakly developed and subject to invasion by cyclones. Nevertheless, the icecaps undoubtedly do have a marked influence on the polar climate as well as on climates of adjacent regions. The cold air masses formed by cooling over the icecaps play an important part in cyclonic activity along the border zone between ice and water and occasionally move into lower latitudes in winter.

The lowest mean annual temperatures on earth are those on the polar icecaps of Greenland and Antarctica. Amundsen-Scott Station at the South Pole recorded a mean temperature of −48.7°C in 1957. Annual ranges are large, although not so striking as at some continental locations in lower latitudes. In Antarctica the annual range is greatest in the interior and least on the coasts, reflecting the effects upon the heat budget of latitude and continentality. McMurdo Sound, on the Ross Sea, has an annual range of 22C° compared with 35C° at Vostok on the interior plateau. Monthly means in summer are normally well below freezing. In addition to providing a cold surface, the ice cover reflects much of the meager solar radiation. Nevertheless, thawing does occur on the ice pack of the arctic near the North Pole. Mushy ice and pools of water are a hazard to explorers travelling over such areas. Antarctica is much colder than the North Polar area in summer. January (Southern Hemisphere summer) temperatures of less than −45°C have been recorded.

Amundsen-Scott 90°S; 2,800 m (1957–60)

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	-28	-38	-55	-58	-56	-59	-58	-60	-59	-51	-39	-28	-49
P (mm)	1	1	T	T	T	T	T	T	T	T	T	1	4
McMurdo Sound, Antarctica 78°S; 20 m													
T (°C)	-5	-8	-16	-23	-23	-24	-26	-27	-22	-19	-9	-4	-17
Vostok II, Antartica 78°S, 107°E; 3,420 m (1958–60)													
T (°C)	-34	-44	-55	-63	-63	-67	-67	-71	-67	-59	-44	-32	-55

Winters in the polar climates are colder still; monthly means range from −20°C to less than −65°C. A temperature of −66°C was recorded at 3,000 m on the Greenland icecap in 1949. Eismitte, also at an elevation of 3,000 m, had a maximum of −3°C and a minimum of −65°C during a two-year period of observations in 1930–31. An extreme minimum of −88.3°C occurred at the Soviet antarctic base, Vostok II (elevation 3,420 m) on August 24, 1960. A feature of temperature trends in the arctic basin, on Greenland, and in Antarctica is a flat curve during the winter months with the minimum at the end of the dark period. This phenomenon has been called the *kernlose* (or coreless) winter. Its occurrence in Antarctica has been explained as follows: After the sun sets at the end of summer the difference between temperatures of the continent and the surrounding oceans brings about instability and initiates cyclones that move maritime air inland to "ventilate" the lower troposphere above a

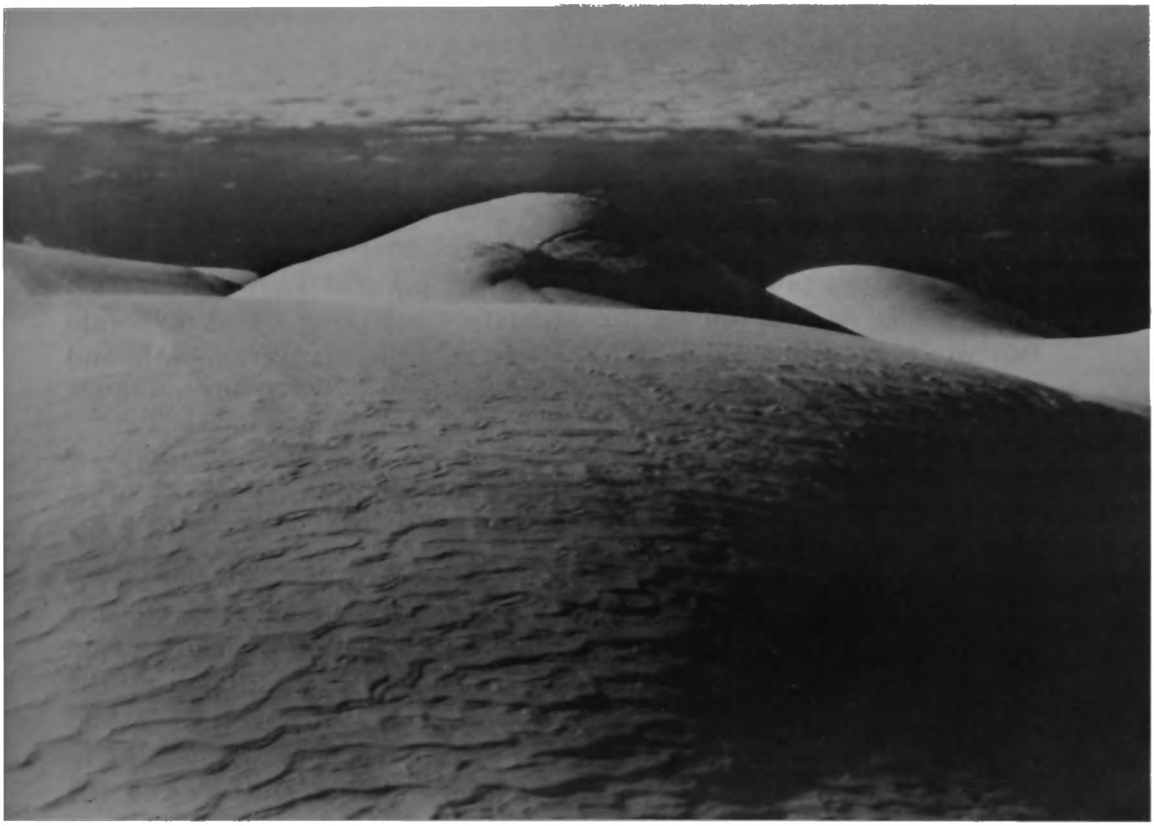


Figure 9.5
Sastrugi on the Antarctic Plateau. (Official U.S. Navy Photograph)

thin layer of cold air, thus slowing the normal decline of surface temperature. By late winter ice covers a much wider belt around Antarctica and reduces the temperature of southward-moving air masses. The result is a secondary drop in temperature to the seasonal minimum.

Diurnal variations of temperature are small throughout the year in polar climates. In summer they decrease generally toward the poles, where the change in altitude of the sun during the day is least.

Cyclonic storms penetrate the polar icecap regions, bringing cloudiness, some snow, and often fierce winds. Precipitation records are fragmentary; possible concentrations of snowfall in a particular season have not been clearly established. Studies of accumulated snow layers on Greenland have confirmed the short-term records of small annual totals. Estimates based on snow and ice profiles in Antarctica indicate annual values ranging from less than 5 cm on the high interior plateau to more than 50 cm on parts of the coastal periphery. When humid air reaches the maritime fringes of the icecaps, formation of frost and rime may contribute to ice accumulations. One principle is certain: as long as an icecap maintains its depth and extent there must be accretions of snow to compensate for losses by sublimation, melting, and break-off around the margins.

Katabatic winds are a marked feature of weather along icecap coasts. Intense radiational cooling forms a shallow layer of cold surface air that flows downslope, sometimes at great speeds and with local gusts that fan the snow into a blizzard. In the interior of Antarctica milder katabatic winds produce unusual snowdrift patterns known as *sastrugi* (Figure 9.5). At Mawson on the Antarctic coast (long. 62°53′ E) southeast winds exceeding 50 knots have been recorded in every month, and gusts have exceeded 100 knots on several occasions.

HIGHLAND CLIMATES

The outstanding feature of highland climates is the great diversity of actual climates which prevail, resulting in large differences over short horizontal distances. The principal regions designated as having highland climates are the main mountain chains and high mountain basins of the middle and low latitudes. These are the Cascade–Sierra Nevada and Rockies in North America, the Andes in South America, the Alps in Europe, the Himalayas and associated ranges as well as Tibet in Asia, the Eastern Highlands of Africa, and mountain backbones in Borneo and New Guinea. They are by no means all of the world's upland areas. Altitude influences climate everywhere, but where other factors overshadow its influence or where the affected areas are small the highlands have been included in other climatic types for the purposes of regional analysis. Near the poles, for example, climates are cold at both low and high elevations, and a detailed examination of the differences is hardly warranted in a world view of climates. The Ahaggar Mountains in the central Sahara have a modified form of the tropical arid climate, but the differences are primarly of the mesoscale.

There is no single, widespread highland climate with a reasonably distinctive combination of elements analogous to, for example, the rainy tropics or the marine climate. Regions with highland climates are mosaics of innumerable microclimates fitted intricately into the pattern of relief and altitude. It would be quite impossible to represent them adequately on an ordinary small-scale map of the world or even of a continent. It is necessary, therefore, to indicate the factors which produce the many small climatic differences and to generalize about the characteristics common to the regions included in this climatic type. The regions considered here as having highland climates are those which have considerable areal extent and which have features distinguishing them from other climatic regions at comparable latitudes. The dominant factors which affect highland climates are: altitude, local relief, and the mountain barrier effect.

The normal lapse rate of temperature entails an average decrease of about 6C° per thousand meters increase of elevation. There are important exceptions to this in actual lapse rates (see Chapter 2, Vertical Distribution of Temperature), but, on the whole, higher altitudes are associated with significantly cooler temperatures. Barometric pressure always decreases with altitude. The atmosphere above areas at high elevations is ordinarily much freer of clouds, dust, smoke, and other nongaseous material, and the air is accordingly more transparent to the passage of both incoming and outgoing radiation.

The influence of local relief is most pronounced in connection with the effectiveness of insolation on slopes having different exposures and with the modification of wind direction and speed. Every variation of slope with respect to the sun's rays produces a different microclimate. In the Northern Hemisphere, the southern slopes of hills and mountains receive more direct insolation, whereas northern slopes are less favored. Many deep valleys and steep northern slopes are exposed to the direct rays of the sun for only a short time each day; in rare cases they may be forever in the shade. Variations in exposure to winds also produce local climates. Wind speeds at high altitudes are generally greater because of less friction with the earth's surface; the higher a site is above the maximum effect of surface friction, the more nearly wind direction conforms to broad patterns of circulation. Winds tend to be gusty in rugged terrain. Locally they may be funnelled through constrictions in canyons or valleys, forced over ridges, or routed around mountains. Under conditions of fair weather, mountain and valley breezes are generated by the relief itself. Thus the direction and speed of wind in mountain valleys may be quite at variance with the air movement aloft or over adjacent plains. Prevailing winds in mountainous terrain often are conditioned more by the trends of valleys than by global or continental circulation systems.

We have treated the barrier effect of mountains earlier in connection with several of the climatic elements. Frictional drag induced by mountains produces "standing waves" that lie perpendicular to wind direction. Stationary lenticular clouds sometimes form along the crest of the wave. The frictional effects may extend to great heights and reduce wind speeds well into the upper troposphere, at the same time generating increased turbulence.

Stable air masses of major proportions can be dammed by a mountain range, creating marked temperature differences between the windward and leeward. In this manner in winter the east-west mountain chains of Europe hold much of the polar air away from the northern Mediterranean coast, the North American Rockies and Cascades protect the Pacific Northwest, and the Himalayas decrease the southward flow of cold air masses from central Asia. Whatever air does overflow these barriers is subject to the stable-air foehn effect and is therefore not so cold when it descends the slopes at the continental margins. (See Chapter 5, Special Air Mass and Storm Effects.) Mountains in the path of moist winds receive considerably more precipitation on the windward side than on the leeward. The accumulations of snow and ice at the higher elevations affect surface albedo, temperature, and air drainage. In general, mountain barriers tend to differentiate the climates of the windward and leeward wherever they lie in belts of prevailing winds of the major circulation systems.

Since highland climates are distributed through a wide range of latitudes, no specific temperature values characterize all such climates. Under clear skies, insolation is intense at high altitudes and there is a greater proportion of the shorter wave lengths (violet and ultraviolet), which do not penetrate well to lower elevations. As a result, one's skin burns or tans more readily. Temperature differences between daylight and darkness and between sunshine and shade are great. One can almost literally "burn on one side and freeze on the other." The same clear air that permits easy transmission of insolation also allows rapid loss of heat by radiation and large diurnal ranges of temperature. Night frosts are common even where daytime temperatures rise well above freezing. A distinctive feature of highland climates in the tropics is a larger diurnal than annual range of temperature. Monthly means and the annual march of temperature are determined primarily by latitudinal variations in insolation and in the transport of air from land or water source regions. Quito, Ecuador, virtually on the equator, has an insignificant annual range appropriate to its latitude (see the accompanying table), but the daily range normally exceeds 10C°. Longs Peak has a much greater annual range (18C°) in response to the variation in insolation through the year.

Quito, Ecuador 0°8'S; 2,811 m

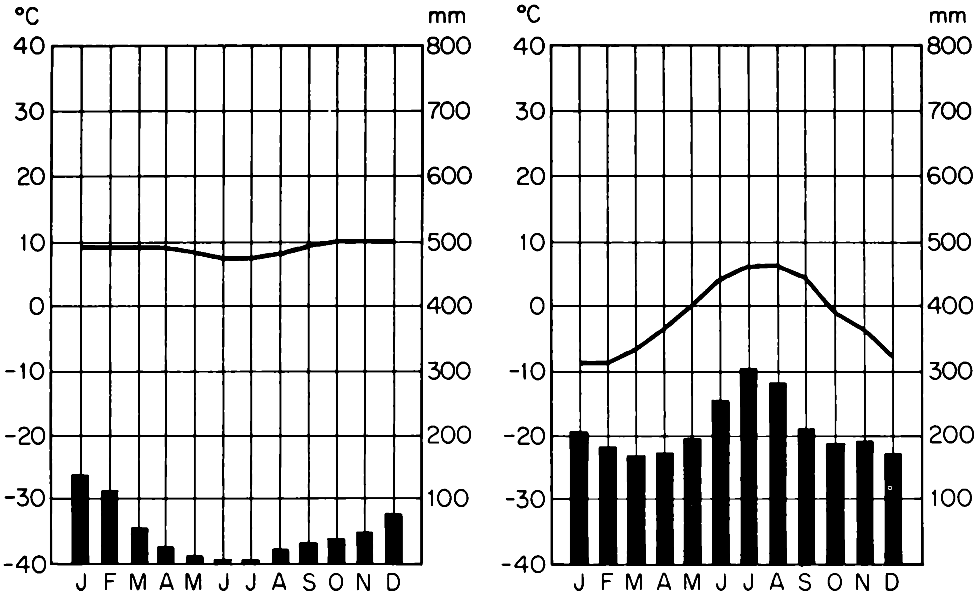
	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
T (°C)	13	13	13	13	13	13	13	13	13	13	13	13	13
P (mm)	119	131	154	185	130	54	20	25	81	134	96	104	1,233

Longs Peak, Colorado 40°N; 2,730 m

T (°C)	−5	−5	−3	1	5	10	13	13	9	4	−1	−5	3
P (mm)	18	30	52	69	60	42	91	57	44	42	22	23	550

Leh, Kashmir 34°N; 3,514 m

T (°C)	−8	−6	0	6	10	14	17	17	13	7	1	−5	6
P (mm)	12	9	12	7	7	4	16	19	12	7	3	8	116



La Paz, Bolivia (1951–60)	16°S; 4103m	Santis, Switzerland	47°N; 2496m
Mean annual temperature	9°C	Mean annual temperature	-2°C
Annual temperature range	3°	Annual temperature range	15°
Mean annual precipitation	555mm	Mean annual precipitation	2488mm

Figure 9.6
Climatic graphs for La Paz and Santis, highland climates.

Whereas temperatures are markedly lower with an increase in elevation, precipitation tends to increase, at least up to altitudes of 3,000 to 5,000 m. Above a zone of maximum precipitation it again decreases, for the orographic effect has already exhausted most of the precipitable moisture. On mountainous trade-wind islands, there is heavy rainfall on low windward slopes, but at higher elevations an upper-air subsidence inversion (the trade-wind inversion) produces arid conditions. As in the case of temperature, no general statement can be made about average amounts of precipitation. Average annual totals are related to the dominant air masses and to the moisture characteristics of prevailing winds. (See Figure 9.6.) Fogginess is generally greater in mountains than on nearby plains or valleys. A great deal of mountain fog is simply clouds formed by one or more of the lifting processes, but radiation fogs are common in calm air on clear nights and in the early morning hours.

With increased elevation there is an ever greater proportion of precipitation in the form of snow, and the snow cover remains for a longer period. The elevation of the lower limit of permanent snow or ice is determined by air temperature, by slope exposure to the sun, and

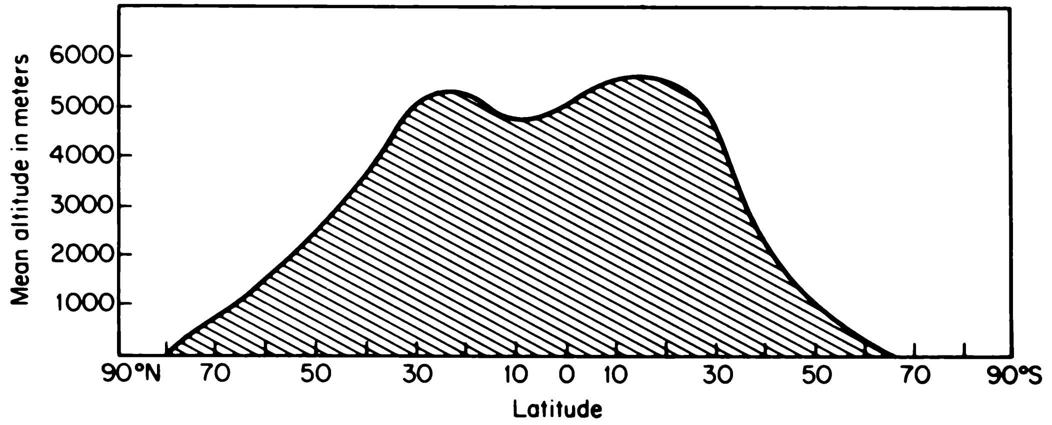


Figure 9.7
Variations of the mean altitude of the permanent snowline with latitude.

by depth of snowfall. Although the snowline is in general higher in the tropics and becomes lower with increasing latitude, it is not highest at the equator but rather in the two drier belts that correspond roughly with the tropical arid and semiarid climates. (See Figure 9.7.) A scanty accumulation of snow obviously can be melted away in a short time under cloudless skies and subsiding air. This also explains the generally higher level of permanent snow on the leeward slopes of mountains and uplands far removed from the oceans. The snowline in the Rockies of Montana is higher than that of the Cascades, which, in turn, is higher than that on the Olympics in the same latitude. Foehn winds account for a more rapid removal of snow on leeward slopes. Other variations from the averages indicated in the figure occur on shaded slopes, where snow will remain much longer than on slopes exposed to the sun. Where there is a winter maximum of precipitation the snowline tends to be lower than in areas at the same latitude with a summer maximum. In any particular location the actual lower limit of snow results from these various factors working together or in opposition.

The causes of precipitation in a highland climate are much the same as elsewhere in the general geographic area in question, except for the intensification by the orographic effect. In the belt of cyclonic storms, for example, the characteristic frontal storms prevail and the annual regime of precipitation is similar to that of the surrounding lowlands, but the annual amount and the total falling in a given storm is likely to be somewhat greater in the highlands.

Additional Readings

HAMILTON, THOMAS D., "Alaskan Temperature Fluctuations and Trends: An Analysis of Recorded Data," *Arctic,* 18, 2 (1965), 105–17.

HARE, F. KENNETH, "The Arctic," *Quart. Journ. Roy. Met. Soc.,* 94, 402 (1968), 439–59.

LAMB, H. H., "On the Frequency of Gales in the Arctic and Antarctic," *Geog. Journ.,* 123 (1957), 287–97.

LEE, R., "The Circulation of the Arctic," *Can. Geog.,* 16 (1960), 1–13.

MACFADDEN, J. D., and R. A. RAGOTZKIE, "Climatological Significance of Albedo in Central Canada," *Journ. Geophys. Res.,* 72 (1967), 1135–43.

ORVIG, S., ed., *Climates of the Polar Regions,* Vol. 14, World Survey of Climatology. New York: American Elsevier Publishing Company, 1970.

RUBIN, MORTON J., "The Antarctic and the Weather," *Scient. Am.,* 207, 3 (1962), 84–94.

SHEAR, JAMES A., "The Polar Marine Climate," *Ann. Assoc. Am. Geogrs.,* 54, 3 (1964), 310–17.

WEXLER, H., "Seasonal and Other Temperature Changes in the Antarctic Atmosphere," *Quart. Journ. Roy. Met. Soc.,* 85, 365 (1959), 196–208.

PART THREE

Weather, Climate, and Life

chapter ten

Climate, Vegetation, and Soils

The world distributions of three elements of man's natural environment show a remarkable coincidence because one, climate, is the dominant factor in shaping the other two, vegetation and soils. Thus climate ultimately affects all forms of life. An earlier discussion of vegetation as an indicator of climatic types treated the influence of climatic elements on plant life in terms of broad correlations. The interrelationship of climate, vegetation, and soils will now be examined in greater detail.

CLIMATIC FACTORS IN PLANT GROWTH

All plants have certain environmental requirements that must be met if they are to thrive. These may be broadly classified under the headings: (1) climatic, (2) physiographic, (3) edaphic, and (4) biotic. The first is of primary concern here, although influences of terrain or relief, soils (the edaphic factor), and the interrelations of plants with other plants and with animals cannot be neglected in the study of plant ecology. Climate does not act alone to set limits to plant growth; rather, it exerts its influence in conjunction with the other factors. Yet its role is always important directly in its effects on plants and indirectly in its influence on edaphic and biotic factors.

The principal direct influences of climate on plants are exerted by precipitation and soil moisture, radiation, humidity, temperature, sunlight, and wind. Variation in one of these factors can change the significance of the others in producing different rates of evapotranspiration and photosynthesis. The moisture factors are the most important over large areas of the earth. Water not only goes into the composition of plant cells; it also serves as a medium for transport of nutrients to growing cells and through evapotranspiration acts as a temperature control. For most land plants the immediate source of moisture is the soil. The amount and availability of soil moisture is not necessarily a simple function of precipitation, but is affected by surface drainage conditions and by the ability of the soil to retain moisture as well as by the losses due to evapotranspiration. Thus swampy areas occur in the midst of deserts and sandy or gravelly soils in rainy climates may be entirely devoid of vegetation. Just as deficient moisture limits plant growth, so excess amounts restrict certain plants by limiting aeration and the oxygen supply in the soil. Excessive soil moisture tends to develop unfavorable soil characteristics and to increase disease damage.

Plants that live in water or in a very moist climate are *hygrophytes;* those adapted to drought are called *xerophytes.* In climates with distinct seasons vegetation must be adapted to alternating cold-hot or wet-dry conditions. Plants with the ability to make the required adjustments are known as *tropophytes.*

The humidity of the air in which plants grow has varying significance depending upon the type of plant as well as upon the soil moisture available to it. Low vapor pressure of the air induces increased losses of moisture through transpiration. Many plants can withstand low humidities so long as their roots are supplied with adequate moisture. The xerophytic vegetation of dry climates is adapted to limited moisture in several ways. Thick waxy bark and leaves inhibit loss by transpiration. Certain plants have root systems that extend deep into the soil as well as over a wide radius to gather moisture from a large volume of soil. Some plants of the desert, notably certain species of cacti, store water during relatively wet periods to be used in dry times. Many desert annuals avoid the prolonged dry spells by passing rapidly through the life cycle from seed to seed after a rain. These are better classified as drought-escaping than drought-adapted.

Whereas moisture provides the *medium* for the processes of plant growth, heat provides the *energy.* Plants can grow only within certain temperature limits, although the limits are not the same for all plants. Certain algae live in hot springs at more than 90°C, and desert lichens withstand 100°C; arctic mosses and lichens survive −70°C. For each species and each variety there is a minimum below which growth is not possible, an optimum at which growth is best, and a maximum beyond which growth stops. Most plants cease growth when the soil temperature drops below about 5°C. If the soil temperature is low the rate of intake of moisture through the roots is decreased and the plant may not be able to replace water lost by transpiration. Freezing temperatures can thus damage the plant cells by producing chemical changes and desiccation. Alternate freezing and thawing are especially damaging. Species differ a great deal in their adaptation to temperature conditions, however, and many plants can endure long periods of below-freezing temperatures although they do not grow. The effect of high temperature is generally to speed up the growth processes. Under natural conditions high temperatures are rarely the direct cause of death in plants. Rather, the increased evapotranspiration induced by the heat causes dehydration of the plant cells. Up to a point this can be forestalled if the moisture supply is adequate. Consequently, the temperature-moisture relationships are quite as important as temperature alone.

The moisture requirements of plants become higher as temperature increases. A given amount of precipitation may result in a moisture deficit in a hot climate, but under cooler temperatures the same amount may exceed potential evapotranspiration and create a moisture surplus. If the demands of evaporation from soil and transpiration from plant surfaces are not met, wilting and eventual death occur. Both evaporation and transpiration are cooling processes that tend to offset the effects of high temperature. Hot winds increase potential water loss and hasten damage to plant tissues; they may be disastrous even when there is abundant soil moisture. Where growing conditions become cooler, as with increasing altitude or latitude, species of plants are successively eliminated from vegetation formations and ultimately all plant life is prohibited regardless of the amount of moisture available in snow, ice, or the frozen soil.

The sunshine factor in plant growth has several aspects. Sunlight has a direct role in photosynthesis and the production of chlorophyll. Without it most plants fail to develop properly, whereas others are adapted to dark or shaded conditions. Light is also a factor in the time required for certain species to flower and produce seed. Ordinarily the amount of light available to growing native plants is sufficient for normal development so that light is not of major importance as a factor affecting geographical distribution of vegetation. However, for an individual plant in a specific environment, light conditions can be critical. Sunshine also is closely related to the temperature factor. Direct radiation combined with atmospheric heat can drive the temperature above the maximum limit permitted by the moisture supply. In moderate intensity it can stimulate plants to their optimum development. The albedo of plant surfaces aids the control of leaf and stem temperatures.

Wind influences vegetation directly by its physical action upon plants and indirectly by accelerating moisture loss. Through convective heat transfer it affects plant temperatures. Plant leaf temperatures are modified by increases in wind at low speeds. As speed increases the proportionate effect is diminished. The rending and tearing action of high-speed or gusty winds is a familiar process. Trees are blown over or stripped of leaves and branches; leaves of bushes are shredded; stems of plants are twisted or broken. Abrasion by windblown sand, gravel, or ice particles can also be quite damaging to plants. These are, however, local effects. There is no direct worldwide correlation of wind belts with the vegetation pattern. Extensive wind damage is common on exposed mountains, where it is combined with the effects of poor soil, low temperatures, and ice or snow cover.

Although mean values of the climatic elements may have broad application in determining the suitability of an area for plant growth, the variations from the normal and the extremes are frequently vital considerations. Occasional droughts, floods, heat waves, or frosts can prove fatal to plants otherwise adapted to the normal conditions. A succession of two or three unfavorable years is particularly devastating. Only those plants which escape the disaster or which are able to withstand it survive. Hence climatic anomalies play a leading part in plant adaptation and natural selection.

Duration of minimum conditions for vegetative growth is another consideration which involves the seasonal distribution of the climatic elements. Whether precipitation is well distributed throughout the year, concentrated in a short season, or is erratic sets broad limits for plant associations. Thus, moisture-loving species of the rainy tropics are not adapted to the hot, dry season of the dry summer subtropics. The duration of temperature conditions is likewise restrictive or permissive relative to the plant association in question. Length of growing season is usually defined as the period between the last killing frost of spring and the first killing frost of autumn. It has its widest usage in connection with crop plants, but short growing seasons set limits to natural vegetation as well. As pointed out in the discussion of climates dominated by polar air masses, the longer daily duration of sunlight in summer at high latitudes intensifies the growing season so that plants are able to concentrate their annual growth into a shorter period than at lower latitudes. Furthermore, plants differ in their susceptibility to frost damage; some are simply hardier than others.

All species of vegetation have climatic optima under which their growth is most efficient. The fact that these optima as well as the extreme climatic limits are different for various plants accounts for the distribution of plant associations on earth. Although the other environmental factors may be favorable they can be negated by adverse climatic conditions.

In summary, heat and moisture are the great determiners of where plants will grow naturally, and the energy and water budgets largely set the world pattern of vegetation.

WORLD DISTRIBUTION OF VEGETATION

The close relation between climate and vegetation is a fundamental relationship in the natural environment. The correlation is in the broad groupings rather than in details and is possible only where undisturbed natural vegetation has reached an approximate equilibrium with the climatic environment. (See Figure 10.1.) Under constantly frozen conditions there is an ice and snow cover or barren rock with no vegetation. Toward progressively warmer climates tundra and taiga prevail, and since even meager precipitation may equal potential evapotranspiration under the low temperatures, moisture is subordinate to temperature as a limiting factor. In warmer regions adequate moisture becomes the dominant factor and there is a succession of desert, steppe, grassland, forest, and rain forest ranging from arid to very wet climates. Each of these vegetation categories has its cool and hot subdivisions. Thus there are

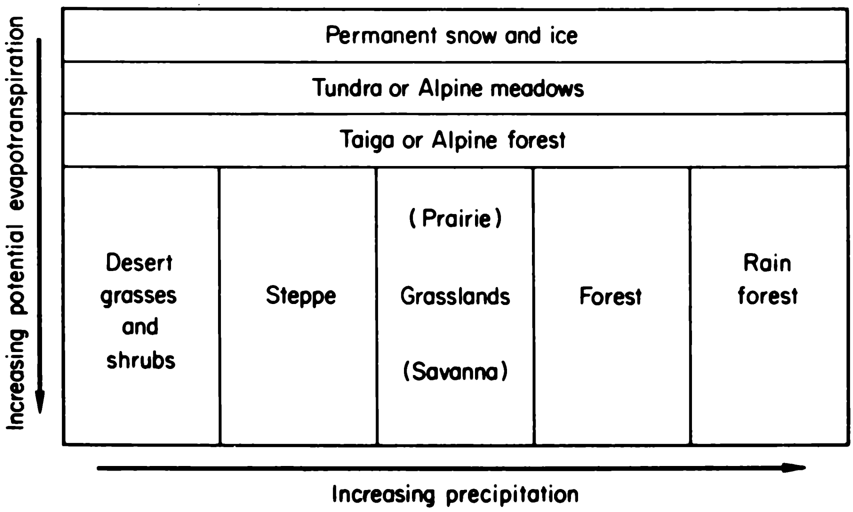


Figure 10.1
Schematic relationship between water budget factors and major world vegetation formations.

cold deserts and hot deserts, prairie grasslands and savanna, subarctic rain forests and tropical rain forests. In each climatic region there is a characteristic vegetation formation consisting of associated plant species which, through the centuries, have become adapted to that climatic region. These dominant forms are the *climax vegetation.* Other species may exist and the climax forms may be locally absent because of soil factors, fire, or human interference, but generally the climax formation reflects climatic influences.

FORESTS

In the rainy tropics the climax vegetation is the tropical rain forest, characterized by many species of tall broadleaved evergreens with spreading, interlocking crowns. Pure stands of a single species are found only rarely. Life seems to be concentrated largely in the treetops. Monkeys and many types of birds are representative animal life. In the densest rain forests, the canopy of overhead growth inhibits the lower plants, but vines push into the treetops in search of light. Epiphytic and parasitic plants also grow in the upper story. Where the tall trees are more widely spaced a dense under story of smaller trees and bush forms fills the openings. On the other hand, the forest floor is relatively clear of vegetation because of the absence of light. Leaves, twigs, and fallen trees are rapidly returned to the soil by insects, bacteria, and the chemical action attending the hot, wet climate. Some species have special requirements which restrict their distribution. The coconut, for example, is found only on or near tropical coasts, and the mangrove is confined to tidal flats. Tropical rain forest, or *selva,* is found in the Amazon Basin, Congo Basin, and on most of the islands off Southeast Asia. Smaller areas occur along several stretches of wet equatorial and tropical coasts. In each region the demands for high temperature and abundant moisture must be met throughout the year, although there are modifications as in the monsoon tropics, where there is a short dry season.

On its drier margins, the tropical rain forest grades into lighter forest associations or scrub and thorn forest. Plant growth is not so luxuriant and the individual species are adapted to lower effective precipitation or seasonal drought. Dense forests are found mostly along watercourses. Typical semideciduous forest has a more open stand of trees than the selva, and sunlight penetrates to the ground to support a thick, impenetrable undergrowth known as jungle. Many of the trees shed their leaves in the cooler dry season and take on drab colors. This seasonal rhythm of vegetation is less evident in the selva. The scrub and thorn forests do not have as many species as either the selva or the semideciduous forests.

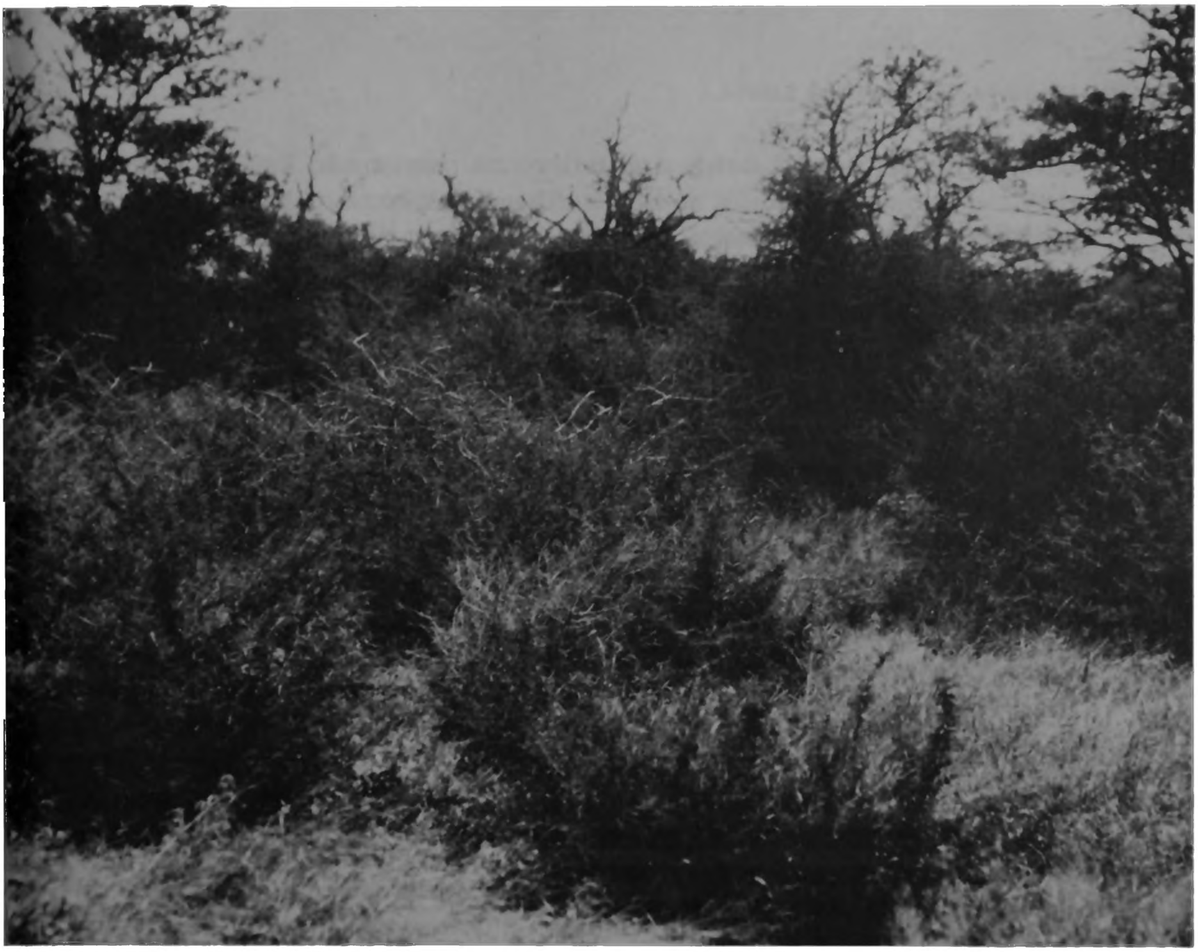


Figure 10.2
Broadleaf tree and shrub association in the low veld of southeastern Africa.
(Photo by author)

(See Figure 10.2.) Trees are relatively small and spaced far enough apart to allow grasses on the forest floor. In some areas, thorns cover the trees as a protective adaptation. Most characteristic of the animal life of the semideciduous and scrub forests are the many insects, notably ants and spiders. The large, so-called "jungle" animals really belong to the adjacent savannas but invade the forests, especially in the dry season, when their normal habitat cannot meet the demands for food.

The mid-latitude forests are divided into three broad classes: Mediterranean scrub, broadleaf and mixed broadleaf-coniferous, and coniferous. Mediterranean scrub forest is made up of broadleaf evergreen scrub types that are adapted to the hot, dry summers and mild, wetter winters. It is also sometimes called the sclerophyllous forest because of the small, thick, waxy leaves and thick bark which enable the plants to resist the desertlike summer. Species vary in the different continents but the forest has the same general open appearance. In southern Europe and California oaks are common trees. The commercially important cork oak of Mediterranean Europe is valuable because of its very thick, spongy bark. Dense thickets of shrubs and bushes are common in Europe, where they are known as *maquis*, and in southern California, where a similar association is called *chaparral*. The Mediterrean scrub forest coincides remarkably wel with the dry summer subtropical climate and therefore has virtually the same worldwide distribution.

The broadleaf and mixed broadleaf-coniferous forests of the middle latitudes flourish where there are moderate temperatures and generally plentiful summer rainfall. They are also known as the temperate or *mesophytic* forests. They have a wealth of species but not as great a variety as in the tropical forests. North America they are found chiefly in the eastern third of the United States, where humid continental climates prevail, but they extend into the humid subtropical climate of the South also to grade into the coniferous belt. Representative trees are the oak, ash, hickory, magnolia, gum, pine, and spruce. Lower forms of natural vegetation as ferns, shrubs, and creping vines. A similar vegetation group was native to a belt stretching from the British Isles through Europe into central U.S.S.R. In eastern Asia the principal regions are eastern Manchuria, Korea, Japan, and eastern China. Throughout the Northen Hemisphere, these forests have been heavily cut over. In China the original forest has ben so completely denuded that its character and extent can be determined only approximately by inference. In the Southern Hemisphere, the mesophytic forests occur in southern Brazil, southern Chile, along the Natal Coast of South Africa, on the southeastern margin of Australia, and in Tasmania and New Zealand. In southern Chile and southwestern New Zealand there are dense subantartic rain forests.

The coniferous forest coincides fairly well with the taiga climate. There are important exceptions, however, on the Pacific Coast of North America, in the Rocky Mountains, New England, and the southeastern United Staes. The later area belongs to the humid subtropical climatic region, and pines are the dominant species. Along the Pacific Coast of Canada and the United States the abundant rainfall has permitted an extratropical rain forest, but conifers predominate. In the taiga the chief specifes are fir and spruce with some pine and larch. Enclaves of deciduous birch and aspen also occur. In general, the density, size, rate of growth, and number of species in the coniferous forest decrease with latitude, and their commerical value likewise diminishes.

GRASSLANDS

The major grassland formations are *savanna*, *steppe*, and *prairie*. Most tropical scrub and semideciduous forests are transitional to the savanna, where tall grasses are dominant but scattered drought-resistant trees and shrubs nevertheless occur. The general appearance is that of an open park with small umbrellalike trees dotted about. The savanna coincides closely with the wet-and-dry tropical climate. Where the wet-and-dry tropics merge into the tropical semiarid climate the short-grass steppe becomes dominant. Outstanding examples of savanna are the *llanos* of the Orinoco Valley and the *campos* of south-central Brazil in South America and the *sudan* in northern Africa. There are also savanna belts in southern Africa and northeastern Australia.

The proportion of trees in savannas varies with the amount of available moisture. Along stream courses, dense growths of trees extend into the savanna to form *galerias* (from the Italian word *galeria,* meaning tunnel). Away from permanent water supplies trees must be adapted to a dry winter season lasting from two to four months. Some are thorny whereas others have thick, leathery leaves which help control moisture losses, and still others are deciduous. The tall, thick-stemmed grasses dry up in winter but revive again with the rainy season. Various large animals feed on the savanna grasses or on tree leaves; these animals are especially numerous in Africa. On other continents there are different animal communities, but the principle of interdependence of elements of the animal kingdom relying on the savanna still pertains. The dry season is a critical period for plants and animals alike.

The steppes consist of a more or less continuous mat of grasses several centimeters in height with shrubs (for example, sagebrush) sometimes dominant. They occur in the mid-latitudes as well as the tropics. Along the dry boundaries xerophytic shrubs are common, and the desert is distinguished more by the open spaces between plants than by the species present. All the great steppe regions are contiguous to desert along at least a part of their dry borders. The tropical steppes have their widest distribution to the north and south of the Sahara, in an irregular strip from Asia Minor eastward to India, on the plateau of South Africa (see Figure 10.3), and around the periphery of the Australian desert. The principal mid-latitude steppes are the North American Great Plains, a belt extending eastward from the Black Sea to Manchuria, and the wetter Patagonia of Argentina. The desert-steppe boundary has been the subject of much study concerning both vegetation and the associated climate. It appears likely that extreme drought years are a factor in determining the boundary zone and resisting the spread of grasses into the desert.

Distinctive grasslands known as prairies border the steppes on their wetter margins in some areas. Tall-grass prairie plants are deep-rooted and reach heights of one to three meters under natural conditions. Short-grass prairies merge into the steppes. The drier edges of the prairies, like the steppes, are probably delimited by dry years. Where they merge into

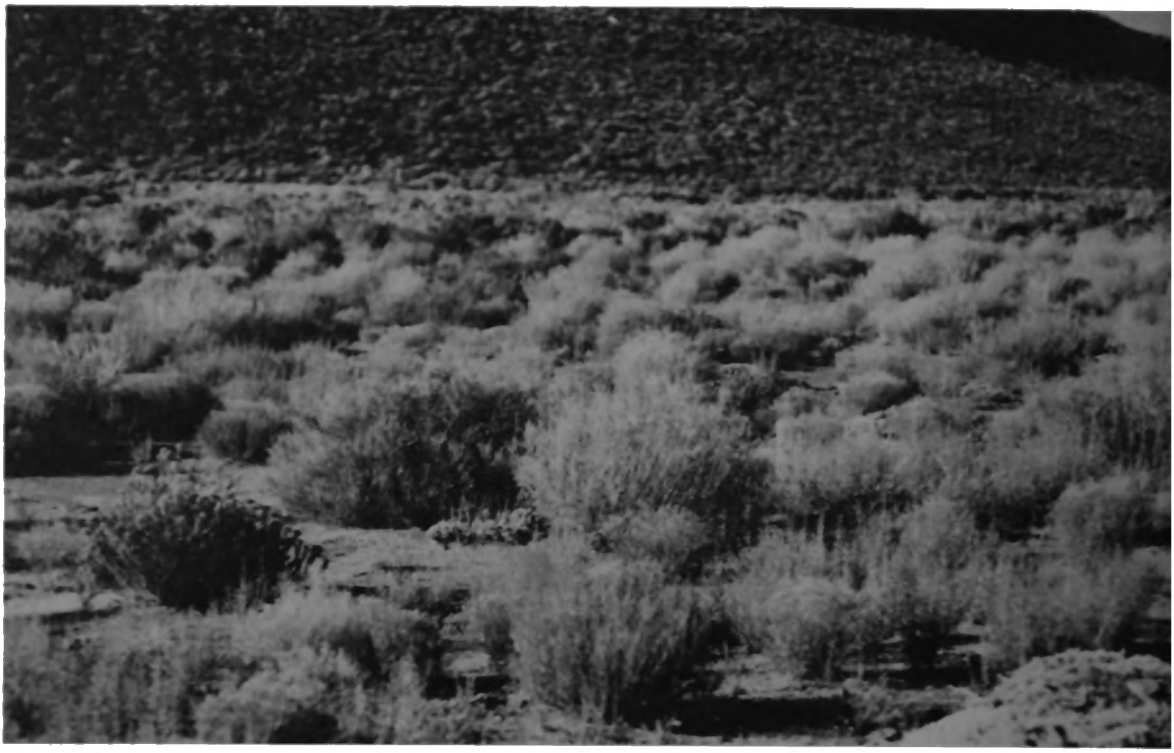


Figure 10.3
Karoo bushes on the semiarid plateau of South Africa. (Photo by author)

forest the transition is usually narrow, although galeria forest may extend along streams. Most of the world's prairies have been plowed up so that few examples of the natural plant associations remain. The principal areas in which prairie grasses were the native vegetation are east of the Great Plains in North America, north of the steppes in U.S.S.R., central Manchuria, the pampas of Argentina, and a small area east of the steppe lands of South Africa. The prairies are associated with both humid continental and humid subtropical climates.

DESERT

To many minds the term *desert* conjures up visions of dry, open expanses entirely devoid of vegetation. Such a concept would limit use of the word to very small areas. From the point of view of plant geography, desert vegetation consists of many species widely spread over the ground and not making a continuous cover. (See Figure 10.4.) The adaptations which characterize the steppe are even more pronounced in the desert. Hard surfaces to prevent water loss, fleshy stems and leaves to store moisture, or extensive root systems are the means by which xerophytic desert plants exist for long, rainless periods. Annuals must complete their life cycle after a chance rain. All species must be adapted to a climate where potential evapotranspiration far exceeds precipitation.

TUNDRA

Tundra as a vegetative region lies between the poleward limits of the taiga and the polar climatic region, which has no vegetation. It represents the forward line of vegetation in its attempt to clothe the earth to the highest latitudes. Low temperatures and permanently frozen subsoil inhibit growth of any but the lower forms of plants. The southern limit of the tundra coincides fairly well with the 10°C isotherm for the warmest month. In the areas with warmer summers the chief vegetation forms are mosses, lichens, sedges, stunted willows, and occasional birches and aspens. Farther poleward there are few, if any, bush forms, and, under the most severe climatic and edaphic conditions, expanses of bare rock, boulder fields, or gravels have practically no vegetation. These latter areas are known as *desert tundra* in contrast to the *grass tundra* and *bush tundra* in the more favorable environments. The annual freezing and thawing of the surface soil causes much heaving and separation, resulting in a distinctive pattern of hummocks or knobs that make the terrain difficult for travel. During summer the surface waters are held in small depressions between these knobs, for the permafrost does not allow down-

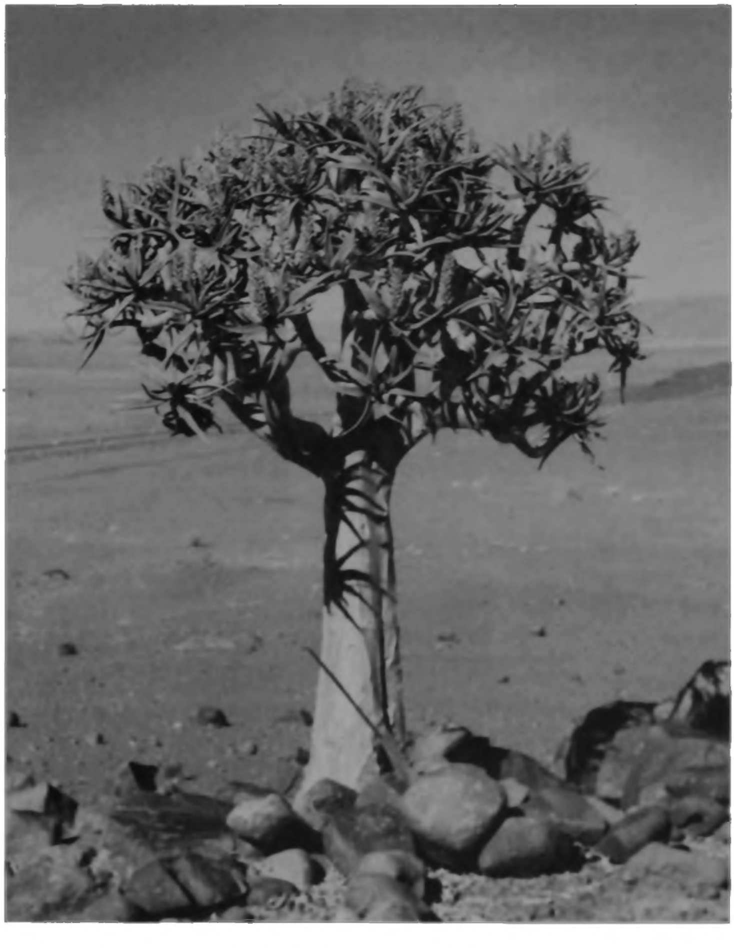


Figure 10.4
Quiver tree (Aloe dichotoma) *in the Namib Desert of Southwest Africa.* (Photo by author)

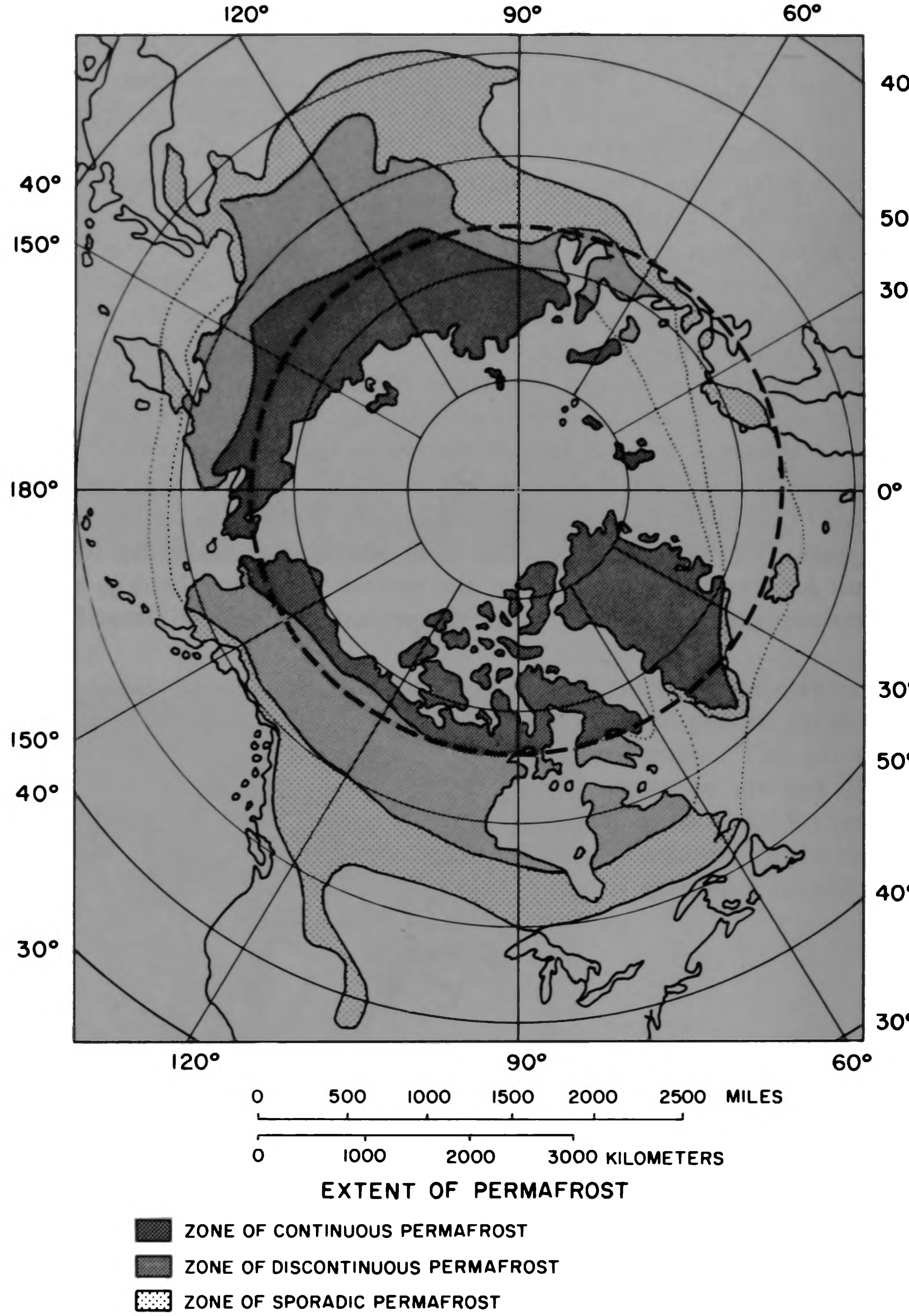

120°
90°
60°
40°
50°
30°
40°
150°
180°
0°
30°
50°
150°
40°
40°
30°
30°
120°
90°
60°
0 500 1000 1500 2000 2500 MILES
0 1000 2000 3000 KILOMETERS
EXTENT OF PERMAFROST
▩ ZONE OF CONTINUOUS PERMAFROST
▨ ZONE OF DISCONTINUOUS PERMAFROST
░ ZONE OF SPORADIC PERMAFROST

Figure 10.5
Distribution of permafrost conditions in the arctic. (From United States Air Force, Geophysics Research Directorate, *Handbook of Geophysics,* Revised Edition, published by The Macmillan Company, 1960)

ward percolation. (See Figure 10.5.) Because there is only slow loss by evaporation, such waters provide breeding places for myriads of mosquitoes. The economic value of tundra plants is limited, but they do furnish forage for animals such as the reindeer, caribou, and the few remaining musk oxen. Fur-bearing animals and birds also inhabit the tundra.

VERTICAL DIFFERENTIATION OF VEGETATION

The vertical zonation of vegetation has already been referred to in connection with the highland climates. The altitudinal range and variety are greatest in the tropics, but, wherever vertical differences in climate occur, corresponding changes in vegetation are likely—unless the climate prohibits plant growth at all altitudes as it does on the polar icecaps. Altitude duplicates, in some respects, the effects of latitude upon the heat balance and upon the types of vegetation; but slope, exposure, cloudiness, wind, and orographic effects also condition the climatic environment of highlands. As highland climates are mosaics of many microclimates with a vertical arrangement, likewise mountain vegetation falls into broad altitudinal belts with many local variations. Thus, the concept of vertical life zones applies regionally, but—in detail—the plant associations vary a great deal. Some of the most striking horizontal variations are on opposite sides of mountains and result from differences in amount of orographic rainfall or in the amount of effective insolation.

Since temperature generally decreases with altitude, whereas precipitation ordinarily becomes greater, the combined result is a rapid increase in the ratio of precipitation to potential evapotranspiration as elevation increases. At the higher levels, the upper limits of plants are more likely to be set by low temperatures than by lack of moisture. (See Figure 10.6.) On the lower slopes (at least in low and middle latitudes), moisture is the critical factor. On some slopes with relatively sparse rainfall the moisture demands of a forest are partly met by fog and clouds which reduce evaporation and deposit quantities of water droplets on the plant surfaces.

Although a given level on a tropical mountain may have annual temperature and precipitation averages like those of a place at a higher latitude, it does not have the same climate. Seasonal rhythm in the climatic elements, especially temperature and precipitation, changes with latitude, and the rhythm of plant growth changes with it.

In the tropical highlands of Latin America, four vertical life zones are commonly recognized in relation to vertical differences in temperature and precipitation effectiveness. The lowest is the *tierra caliente*

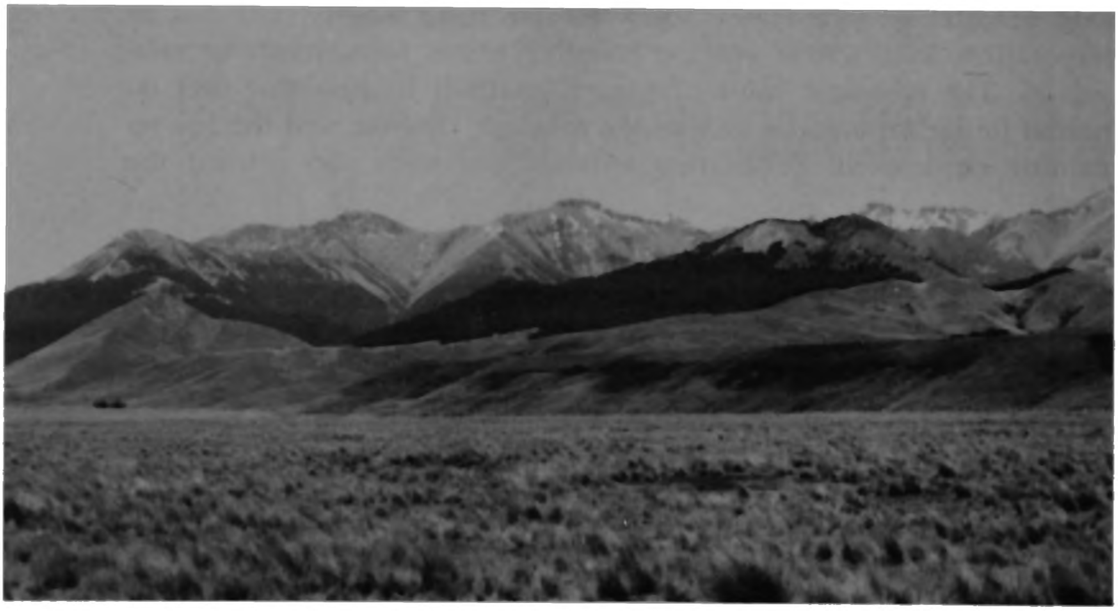


Figure 10.6
Vertical zonation of vegetation in southern New Zealand. A belt of evergreen beech forest lies between the tussock grassland (foreground) and the alpine plants and rock outcrops of the higher slopes. (Photo by author)

(literally, hot land), extending from sea level up to 750 or 1,000 m and having vegetation typical of the tropical lowlands. Annual temperature averages are generally above 25°C. Above the *tierra caliente,* the *tierra templada* (temperate land) rises to elevations of 2,000 to 2,200 m and embraces a modified form of tropical forest. Temperatures range between 25°C and 18°C on the average. Higher still is the *tierra fria* (cold land) from 2,000 to 3,700 m. There the forest becomes less dense and grades into bushy types. The upper limit of the *tierra fria* coincides with the upper limit of cultivated crops, which is at about the 12°C annual isotherm. The *tierra helada* (frost land) extends on upward to the permanent snowfield. It is a zone of alpine meadows, whose character depends upon the amount and seasonal distribution of precipitation. Where there is adequate moisture throughout the year the grasses are relatively dense and perpetually green. Farther from the equator, there is a short dry season and the alpine meadows are dry for a part of the year. Toward the regions of subsidence in the subtropic highs, conditions become drier and the associated vegetation grades into thorny and desert forms.

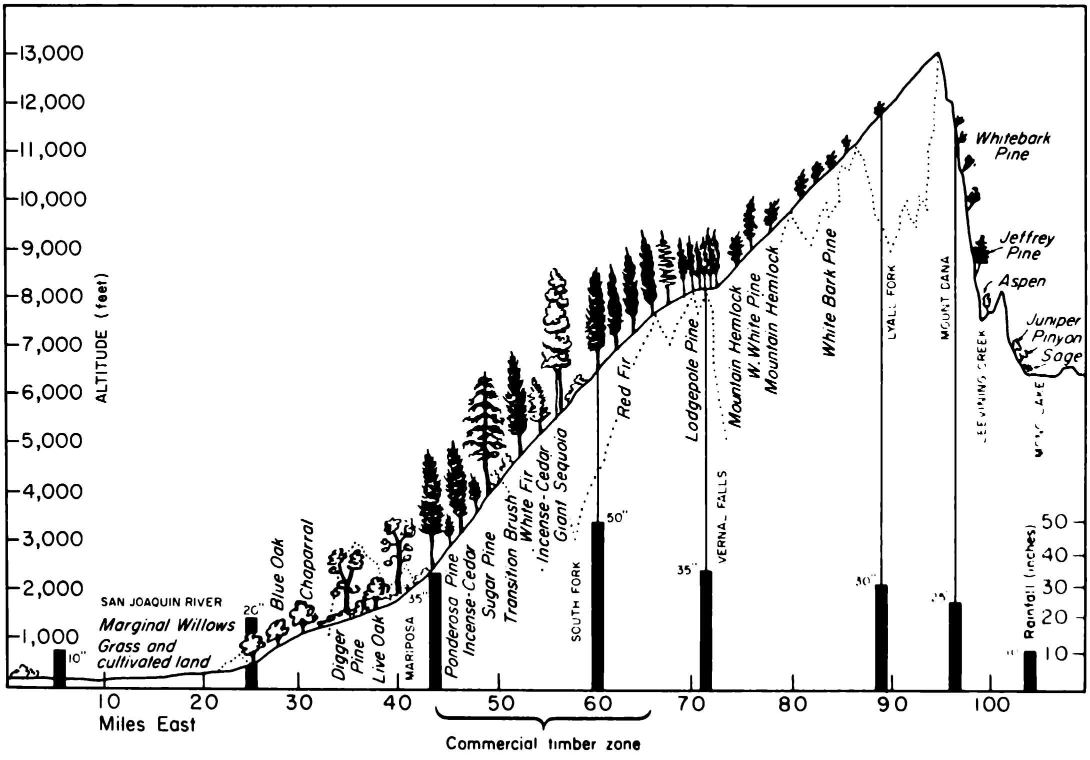
In the middle latitudes, the *tierra caliente* is absent and there is actually nothing akin to the *tierra templada.* Altitudinal limits to vegetative types are lower, but moisture conditions still have much influence on both horizontal and vertical distribution of plants. Figure 10.7 illustrates with a specific example the vertical zonation of vegetation in the Central Sierra Nevada region of California.

CLIMATE AS A FACTOR IN SOIL FORMATION

Five natural factors influence the formation of soils: (1) climate, (2) plant and animal life, (3) relief, (4) parent material, and (5) time. Of these climate is the most active. Soils are affected by climate directly throughout their evolution from parent rock to their current state of development. Their character is shaped indirectly by climate acting through vegetation and animal life. Many of the effects of varied relief are ultimately climatic as a result of local microclimates produced by exposure, slope, drainage, or altitudinal differences.

Soil is a layer of earth material which undergoes constant change and development; it is not merely an inert mass of finely divided rock. The dynamic processes which form soils are of three broad classes: physical, chemical, and biological. Physical breakup of parent rock, termed *disintegration,* is induced climatically through the action of rain splash, rainfall and snowmelt runoff, glacier movements, freezing and

Figure 10.7
Vertical zonation of vegetation in the Central Sierra Nevada. (From *Trees, 1949 Yearbook of Agriculture*)



thawing, and abrasion or transport by wind. Chemical weathering, or *decomposition,* results from chemical interaction of various compounds, and it takes place more rapidly under warm, humid conditions than in cold or dry climates. Rocks softened by chemical action are more easily worn away by physical processes and become more suitable for plant growth. Once the weathered parent material begins to support life, plants, bacteria, worms, and so on speed up the chemical and physical changes and add organic matter. Within certain limits, chemical and biological activity increase with increased temperature and moisture; neither is possible without water. Downward transport of colloidal solids is called *eluviation,* whereas *leaching* is the term given to removal of mineral or organic compounds in solution. Both tend to impoverish the top layers of the soil. On the other hand, where there is lack of moisture there is accordingly limited downward percolation and capillary water tends to leave concentrations of dissolved salts at the surface when it evaporates. The salt flats and alkaline soils of arid climates are examples of soils formed in this way. Broadly speaking, the soils of deserts and steppe grasslands are formed where evaporation exceeds precipitation, whereas forest soils are associated with the reverse conditions. Thus total precipitation is not the sole control of moisture in soil formation. Cloudbursts tend to puddle the surface soil and seal it off to downward water movement so that there is excessive surface runoff. Gentle rains have time to soak into the soil and are more significant in soil-forming processes requiring moisture. Obviously the degree of permeability of the soil will also affect water percolation. Whether the rate of percolation is rapid, as in sands and gravels, or slow, as in clays, greatly affects the usefulness of precipitation. Wind is a drying agent as well as an agent of erosion. By increasing evaporation from the soil it increases the water need of plants.

In the processes of soil formation the climate within the soil is quite as important as that of the air above. Temperature changes are conducted downward slowly in soil so that at a depth of two or three feet diurnal variations are not experienced. Seasonal variation lags considerably at the greater depths, and at about 16 m below the surface there is little or no seasonal temperature change. The structure of the soil and its moisture content help to determine its conductivity; dry, loose soil conducts heat slowly, whereas wet, compact soil and solid rock are much better conductors. Air, a very poor conductor, reduces heat transfer in porous soils. Similarly, snow or a mantle of plant litter inhibits heat conduction. Rain water and snowmelt tend to carry heat downward as they penetrate into soil pores, however.

Except in the very dry deserts, the air in the soil below the immedi-

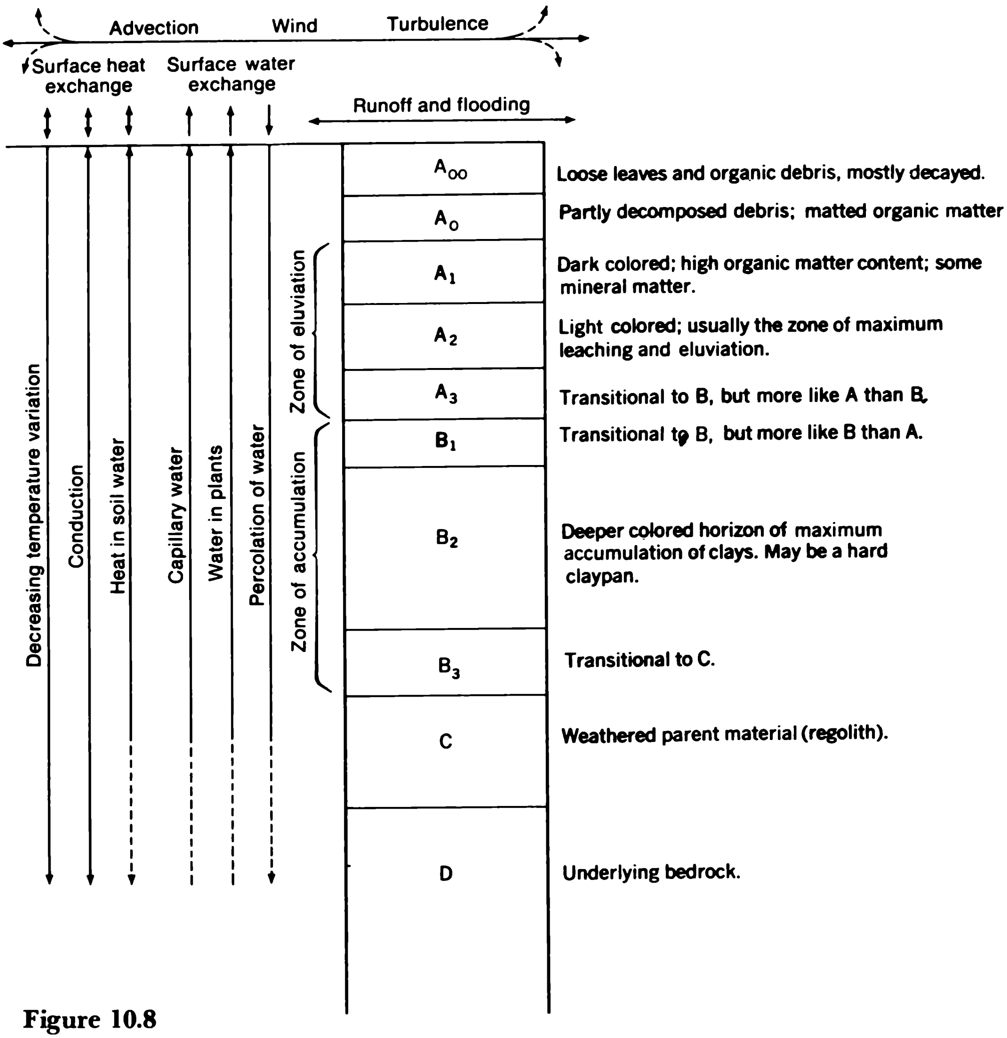

Advection
Wind
Turbulence
Surface heat exchange
Surface water exchange
Runoff and flooding
Decreasing temperature variation
Conduction
Heat in soil water
Capillary water
Water in plants
Percolation of water
Zone of eluviation
Zone of accumulation
A_{00}
A_0
A_1
A_2
A_3
B_1
B_2
B_3
C
D
Loose leaves and organic debris, mostly decayed.
Partly decomposed debris; matted organic matter
Dark colored; high organic matter content; some mineral matter.
Light colored; usually the zone of maximum leaching and eluviation.
Transitional to B, but more like A than B.
Transitional to B, but more like B than A.
Deeper colored horizon of maximum accumulation of clays. May be a hard claypan.
Transitional to C.
Weathered parent material (regolith).
Underlying bedrock.

Figure 10.8
Hypothetical soil profile and major exchanges of heat and moisture.

ate surface is ordinarily saturated, a fact of considerable importance in both chemical and biological processes so long as temperatures are favorable. The relative humidity within the soil is sometimes determined by the relative humidity of the atmosphere. If the average relative humidity of the air is close to saturation (90 to 100 percent), the soil climate is moister than in regions of low humidity even though rainfall may be less.

Assuming that the weathered parent material is not eroded away (as is likely on a steep slope), the various soil-forming processes combine to produce layers, or *horizons,* in the soil and over a period of time a distinctive *profile* is developed. The soil profile is the most useful feature for soil identification and classification. The characteristics of color, texture, structure (arrangement of soil particles), and the chemical and organic matter content vary downward through the profile and afford a basis for study of the origin of the soil. Figure 10.8 is an idealized soil profile. Not every soil should be expected to have all of the horizons, but it will have some of them.

SOIL CLASSIFICATION

The degree to which characteristics of soil profiles are related to the major soil-forming factors is a basis for a genetic classification of soils. In the nineteenth century Russian pedologists developed a classification which is still widely used in various modified forms. A modification employed in the United States and several other countries incorporates three broad soil orders: zonal, intrazonal, and azonal.

Zonal soils have well-differentiated horizons resulting from strong climatic and biological influence. Their distribution shows a close correlation with climate.

Intrazonal soils reflect the dominance of relief or parent material in their formation. Microclimatic effects or soil climate induced by relief may be expressed in their profiles. For example, bog soils are often formed in depressions where there is excessive soil moisture, even though the atmospheric climate is relatively dry.

Azonal soils have poorly developed profiles, usually reflecting the

TABLE 10.1
Bioclimatic Relationships of the Ten Soil Orders (United States Comprehensive System)*

Soil Order	*Moisture Conditions*	*Temperature*	*Predominant Vegetation*
1. Entisol	No climatic or biotic correlations		(Immature soils)
2. Vertisol	Subhumid to arid	Temperate to tropical	Grassland
3. Inceptisol	Humid	Arctic or alpine to tropical	Forest
4. Aridisol	Arid	Temperate to tropical	Desert
5. Mollisol	Semiarid to subhumid	Subarctic and alpine to tropical	Steppe and prairie
6. Spodosol	Humid	Subarctic to tropical	Coniferous forest, savanna, tropical forest
7. Alfisol	Subhumid to humid	Subarctic to tropical	Coniferous and deciduous broadleaf forest
8. Ultisol	Humid	Temperate to tropical	Forest, savanna, swamp
9. Oxisol	Subhumid to very humid	Subtropical; tropical	Savanna, forest, rain forest
10. Histosol	No climatic or biotic correlations		.(Organic soils)

*Based on *Soil Classification, A Comprehensive System* (Washington: Soil Conservation Service, 1960).

short time during which they have undergone the soil-forming processes. Recent deposits of alluvium or loess, sand dunes, and partially weathered bedrock are examples of azonal soils.

Since the middle of the twentieth century soil classification has become more empirical, not only to achieve greater objectivity but also to facilitate application in a widening range of soil problems that extend beyond plant geography and agriculture. In 1951 the Soil Survey Staff of the U.S. Soil Conservation Service undertook the development of a new comprehensive soil classification based on soil properties. The resulting system divides all soils into 10 orders (Table 10.1). Inasmuch as the primary criteria for the orders and their subdivisions are soil characteristics rather than genetic factors the correlation with climate is limited. The classification nevertheless permits selection of differentiating properties that reflect soil-forming processes.

SPATIAL PATTERNS OF SOILS

To the extent that climatic conditions influence the development and properties of soils major world soil regions correspond in a general way to the world climatic regions and, therefore, to the spatial patterns of vegetation. A genetic classification obviously enhances the degree of correspondence for soil categories that are defined in terms of climatic or biological factors.

In spite of the dangers that attend prejudgment of soil genesis, we can make some useful generalizations about soil-forming processes in different climates (see Figure 10.9). Under permanent snow and ice soils cannot develop. In the tundra poor vertical drainage and low temperatures result in distinctive shallow, acidic, and often peaty soils of little agricultural value. Leaching of colloids and bases proceeds at a greater rate under the higher temperatures, longer summers, and moisture surpluses of the taiga climate; this forms characteristic cold-forest soils that have a light-colored (ashy gray) A_2 horizon. In warmer forest regions leaching is less significant, and the resulting profiles are neither as acidic nor as well developed. Under deciduous and mixed forests in the humid continental climates and in parts of the marine and dry summer subtropical climates the soils are not only less acidic but also lack both the concentrated horizon of organic matter at the surface and the gray A_2 horizon. (See Figure 10.10.) Where still warmer, humid conditions prevail decomposition of organic matter and leaching of bases and other soluble materials takes place even more rapidly to produce acidic soils whose *B* horizons are stained by iron compounds. Such soils are ap-

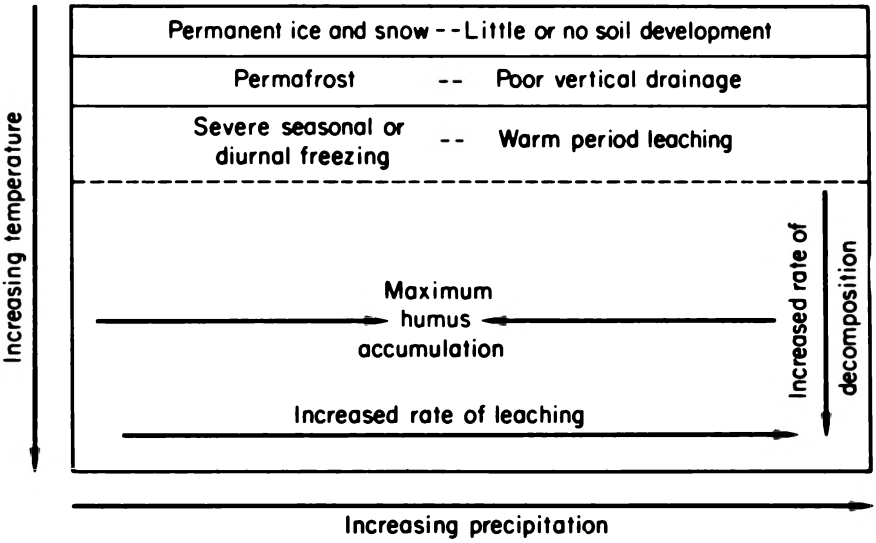
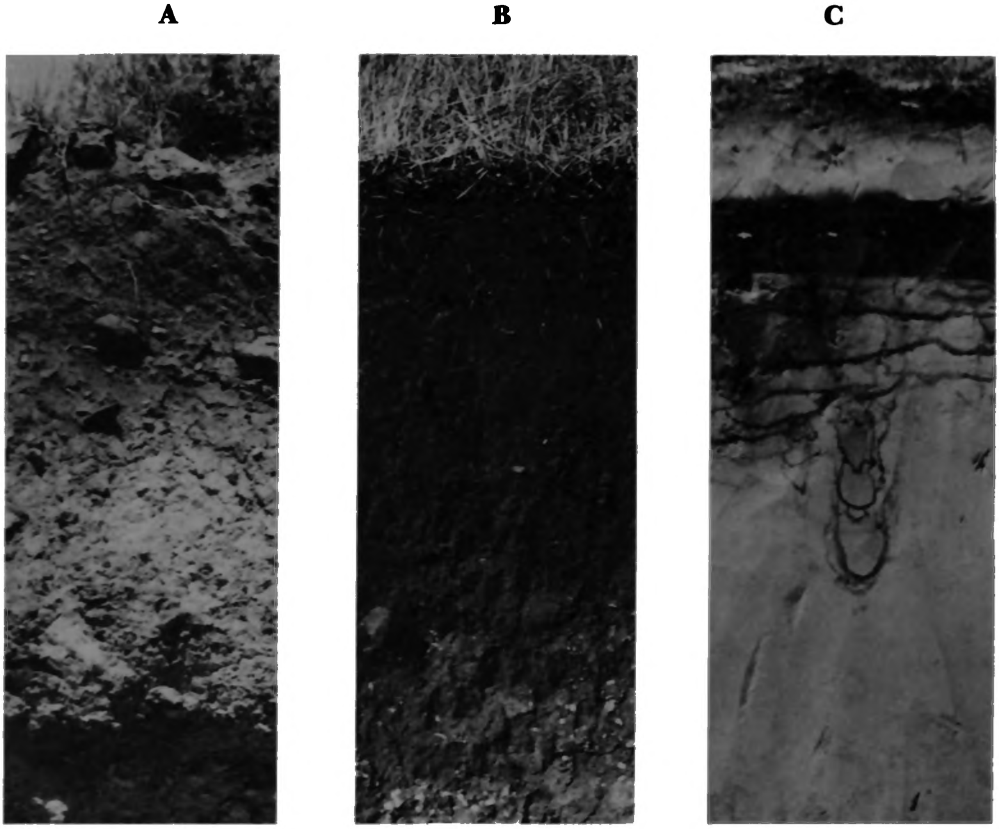

Permanent ice and snow--Little or no soil development
Permafrost -- Poor vertical drainage
Severe seasonal or diurnal freezing -- Warm period leaching
Maximum humus accumulation
Increased rate of leaching
Increased rate of decomposition
Increasing temperature
Increasing precipitation

Figure 10.9
Schematic relationship of major soil-forming processes to temperature and precipitation.

Figure 10.10
Contrasting soil profiles. Each profile represents a depth of one meter. **A**, *dry climate profile in stony parent material.* **B**, *subhumid climate and dense grass cover develop a deep* **A** *horizon darkened by humus.* **C**, *an organic layer, covered by silt deposits, is leached by percolating water in a humid climate.* (Soil Conservation Service—USDA)



proximately coextensive with the humid subtropics and are transitional to those of the tropical humid climates. They are often classed with certain tropical soils because the processes that accompany high temperatures and rainfall account for several of their characteristics.

In the humid tropics bacterial activity and rapid chemical action, combined with excessive leaching, keep humus to a bare minimum. Deep profiles and an abundance of iron and aluminum compounds typify soils of the rainy tropics, monsoon tropics, and wet-and-dry tropics. The native vegetation of subhumid climates is primarily grassland. Sustained surpluses of moisture are uncommon, and the processes of leaching and eluviation are slowed accordingly. The abundance of organic matter from grass roots and decayed leaves makes the *A* horizon much darker than that of forest soils. The change is gradual from the soils of humid areas, which are leached and eluviated, to those of drier regions, which experience far less percolation of salts and colloids. Calcium carbonate accumulates at increasingly high levels in the profile; acidity decreases as precipitation fails to meet the needs of potential evapotranspiration. On the steppe grasslands of semiarid climates alkaline compounds accumulate near the surface, and since the amount of organic matter diminishes progressively as conditions become drier the soils are lighter in color and have shallower *A* horizons. Plants that tolerate large amounts of base minerals tend to aid the concentration of calcium salts near the surface.

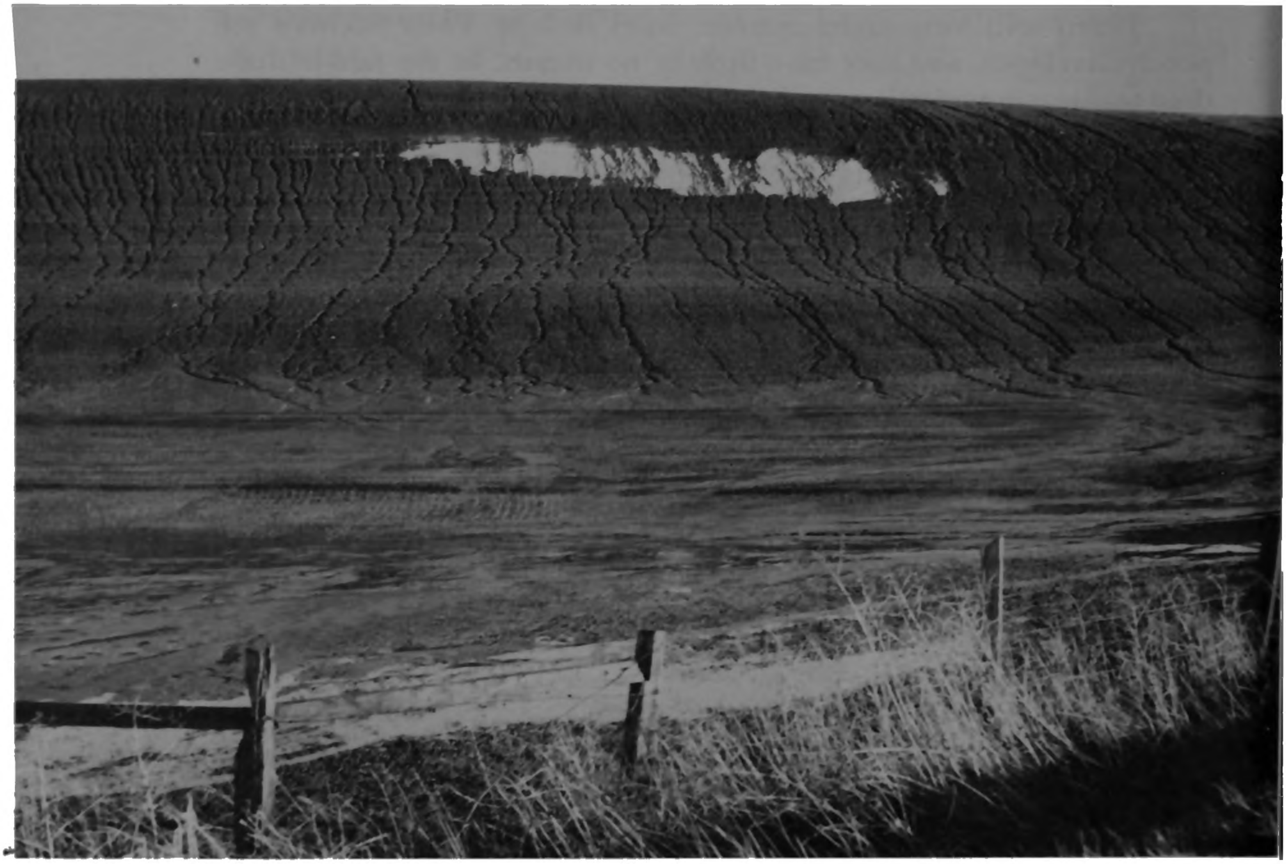
Desert soils form under extreme water deficits. Their horizons are poorly developed, and they have little or no humus. In the mid-latitude deserts they are typically gray in color; in the tropical deserts a reddish color resulting from certain iron compounds is more typical. The accumulation of soluble salts may be exposed at the surface, especially in poorly drained depressions or flat basins, where surface waters evaporate to leave saline deposits. These *evaporites* are valuable clues to past climates.

The effects of altitude and relief on climate are reflected in the soils of highland regions. Just as highland climates are mosaics of many local climates, so soils of mountains and valleys vary greatly over short distances. Their complex spatial patterns make it impossible to represent them on a small-scale map. Many are shallow and immature because of steep sites and rapid erosional processes. Others form under unique local conditions of climate induced by relief and drainage. Only on gentle slopes, on flat plateaus, or in broad valleys are soils having well-developed profiles likely to be extensive. The altitudinal zonation of mountain soils coincides broadly with highland vegetation—the effect of altitude approximates that of latitude. Alpine soils, for example, are roughly analogous to tundra soils. Acting directly and through the dominant vegetation associations, the changing water budget produces a succession of soils that correspond to the life zones of plants.

SOIL EROSION

We have already seen that the natural processes of erosion are active in the development of soils and that they are closely associated with climate. Our present concern will be with accelerated erosion, which tends to destroy soils that are more or less in harmony with natural conditions. The immediate climate-connected causes of accelerated erosion are the same as those which weather and transport parent rock in the early stages of soil formation; consequently, the distinction between "natural" and "accelerated" erosion is not always clear. But in one way or another, man is usually an indirect agent in accelerated erosion. Where there is excessive precipitation and hence a great potential danger of erosion due to surface runoff, a dense vegetation cover ordinarily exists to inhibit soil erosion. In arid climates vegetation is sparse, but running water is not

Figure 10.11
Rainfall and snowmelt rapidly erode fine-textured loess soils on steep slopes of the Palouse region in eastern Washington. (Soil Conservation Service —USDA)



such a great threat to the soil. Thus, there is a semblance of equilibrium between the forces that form the soil and those that would erode it. Man's contribution is essentially a negative one. Cutting forests, cultivation, grazing of herds, or burning of the plant cover all serve to upset the natural equilibrium and to accelerate erosion. Once the protective cover of vegetation and the organic matter in the topsoil are reduced, erosion proceeds at a more rapid rate. Cultural practices in land use have all too rarely provided for a compensating acceleration of soil-forming processes. Many soil-building practices are known, but they are beyond the scope of this book.

The world distribution of actual accelerated erosion, being in part the result of land-use practices, does not show a close correlation with the patterns of climate and vegetation. Rather it is the potential erosion, or what might be called the "erosion hazard," which is closely related to climatic conditions and to the associated soils. The fact that some soils erode more easily than others is traceable, in part, to the effect of climate upon their formation. (See Figure 10.11.) In the regions of heavy rainfall, running water readily attacks exposed soils. In the humid tropical climates the soils are generally deficient in humus so that, although the clay subsoils are quite resistant to erosion, the topsoil washes away easily. Where the rainfall is seasonal, where the soil is frozen for a part of the year, or where there is a snow cover for an extended period, the forms and intensity of erosion on exposed soils are altered but may be none the less disastrous. When frozen soil begins to thaw, melting usually takes place at the top first, and, until thawing is complete, vertical drainage of water is impeded. Under these conditions, moving surface water can rapidly carry away layers of mud. Snow affords a protective cover until it melts; thereafter the melt water from heavy snowpacks becomes a particularly active agent in sheet and rill erosion. The intensity of rainfall, as well as the amount, is important in determining the degree of erosion due to runoff. Thundershower precipitation does not penetrate the soil as well as light continuous rains, and the resulting runoff and erosion are much greater, especially in dry lands. One cloudburst can do far more damage than several months of gentle rains. Similarly, the concentration of a given amount of rainfall in one season creates an erosion hazard greater than that of an area with the same rainfall total more evenly distributed through the year. Heavy rains which break a period of drought commonly cause considerable loss of soil, for vegetative cover is at a minimum and the loose topsoil is easily removed.

Soils in regions of arid and semiarid climates are the most susceptible to the ravages of wind erosion. Soil water not only adds to the weight of the particles but is a cementing agent which reduces the tendency of soil to be blown about. When exposed to the wind, soils

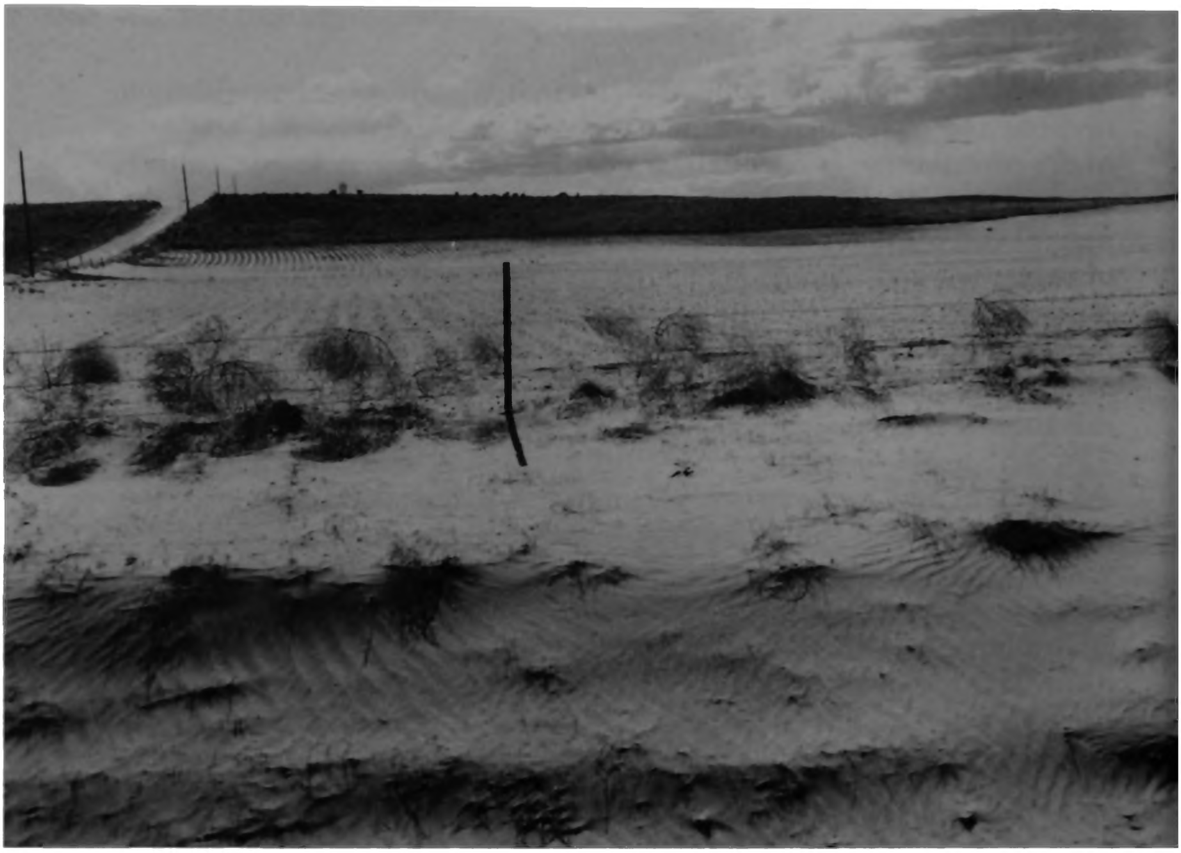


Figure 10.12
Topsoil eroded by wind in the semiarid climate of western Nebraska. (Soil Conservation Service—USDA)

lacking in moisture are readily blown if the particles are small enough. Unfortunately, wind, like water, has a sorting action. It picks up and removes the finer particles which are the basis of soil fertility and leaves the heavier material in drifts or dunes. (See Figure 10.12.) Its earth-moving capacity is not as closely related to slopes as in the case of erosion by water. In the drought years of the 1930s in the "Dust Bowl" of the United States, the loose topsoil was completely removed from large areas, leaving only the compact clays of the subsoil. Semiarid grasslands are especially critical regions for wind erosion when excessive disturbance of the plant cover (through overgrazing or overcultivation) coincides with unusually dry years.

Temperature has its maximum effect on accelerated erosion indirectly through its influence on plant cover and the weathering processes. Freezing and thawing directly alter the structure of the soil and thus make it more susceptible to the action of wind or running water. When frozen for a continuous period, the soil is largely spared from erosion. The soils of the tundra and taiga are singularly free from accelerated erosion, mostly because of the lack of cultural disturbances but partly owing to the long periods of freezing temperatures.

Additional Readings

CRITCHFIELD, HOWARD J., "A New Comprehensive Soil Classification: Its Implications for Geography," *Geog. Rev. of India,* 25, 3 (1963), 153–63.

GATES, DAVID M., "The Energy Environment in Which We Live," *Am. Scientist,* 51, 3 (1963), 327–48.

HADLOW, LEONARD, *Climate, Vegetation, and Man.* New York: Philosophical Library, 1953.

HOLDRIDGE, LESLIE R., "Determination of World Plant Formations from Simple Climatic Data," *Science,* 105, 2727 (1947), 367–68.

LARSEN, JAMES A., "Vegetational Relationships with Air Mass Frequencies: Boreal Forest and Tundra," *Arctic,* 24, 3 (1971), 177–94.

MAJOR, JACK, "A Climatic Index of Vascular Plant Activity," *Ecology,* 44, 3 (1963), 485–98.

MATHER, JOHN R., and GARY A. YOSHIOKA, "The Role of Climate in the Distribution of Vegetation," *Ann. Assoc. Am. Geogrs.,* 58, 1 (1968), 29–41.

RICHARDS, P. W., "Plant Life and Tropical Climate," in *Biometeorology,* ed. S. W. Tromp. London: Pergamon Press, 1962.

SIMONSON, R. W., "Soil Classification in the United States," *Science,* 137, 3535 (1962), 1027–34.

TOSI, JOSEPH A., JR., "Climatic Control of Terrestrial Ecosystems: A Report on the Holdridge Model," *Econ. Geog.,* 40, 2 (1964), 173–81.

chapter eleven

Climate and Water Resources

Of all the earth's resources none is so fundamental to life as water. The properties of water in its three physical states make it by far the most useful of compounds. We can breathe it, drink it, bathe in it, travel on it, or see beauty in it in one or more of its forms. It is a raw material, source of power, waste disposal agent, solvent, medium for heat transfer, or coolant as the needs of modern technology may require. The high specific heat of water, its ability to exist in gaseous, liquid, or solid forms under natural conditions, and its capacity for storing and releasing latent heat with changes of state give it immense influence over atmospheric processes. The availability of water at different times and places is ultimately related to weather and climate. The restless atmosphere is the most active agent in the constant redistribution of water on the earth's surface—a fact that becomes all the more striking when we realize that only a minute fraction of 1 percent of the earth's water is contained in the atmosphere at any one time. If all the atmospheric moisture were precipitated it would create a layer averaging only about 2.5 cm deep over the entire globe. The seas and oceans contain about 97 percent of all the earth's water; 2 percent is in the snowfields and glaciers; and fresh-water bodies and ground water account for nearly 1 percent.

THE HYDROLOGIC CYCLE

The constant circulation of water from oceans to air and back again to the oceans, entailing residence of varying duration in life forms, fresh-water bodies, ice accumulations, or as ground water, generates the *hydrologic cycle,* or simply the water cycle. It has no "beginning" or "end." Figure 11.1 shows that it is an intricate combination of evaporation, transpiration, air mass movements, condensation, precipitation, runoff, and ground-water movements. While the greater part of moisture which eventually falls as precipitation comes from the oceans, some water takes a short cut in the cycle and enters the air directly through evaporation and transpiration from soil and vegetation. But only a small part of precipitation over land can be traced back to evapotranspiration from the same area. Precipitation over the oceans is another form of hydrologic short cut. Other water completes the cycle with fresh-water bodies taking on the function of the oceans. A slowdown in the cycle for various periods is brought about when plants and animals use water for cell building, when icecaps and snowfields detain it in a solid state, when chemical action incorporates it in other compounds, or when ground water is trapped. But, sooner or later, most terrestrial water appears elsewhere in the hydrologic cycle.

The part of the hydrologic cycle concerned with atmospheric moisture has been treated in Chapter 3. The processes of evaporation, condensation, and precipitation are essentially climatic; their functions in the water cycle would be simple indeed if it were not for the constant motion in the atmosphere. Thus, the patterns of atmospheric circulation have a bearing upon the regional manifestations of the hydrologic cycle. Dry air masses assume a part of the task of removing water from the continents. When they warm as a result of subsidence or movement toward the equator, they are especially effective in speeding up evapotranspiration. Their moisture content may also be increased by mixing with maritime air or by evaporation of rain falling from overriding maritime air. Water gained in this manner is precipitated again in storms over land or sea, or the air masses are further modified over oceans to become maritime. Maritime air masses carry water onto the land. Wherever they undergo the lifting processes—convection, orographic ascent, or convergence—precipitation is likely to result. The hydrologic cycle requires heat as well as moisture. Where heat and moisture are abundant, as in the rainy tropics, the cycle is active. In dry climates an essential part of the cycle, atmospheric moisture, is lacking. In very cold climates, the energy to operate the cycle is limited. Were it not for the encroachment of warm maritime air masses into high latitudes, there would be far less

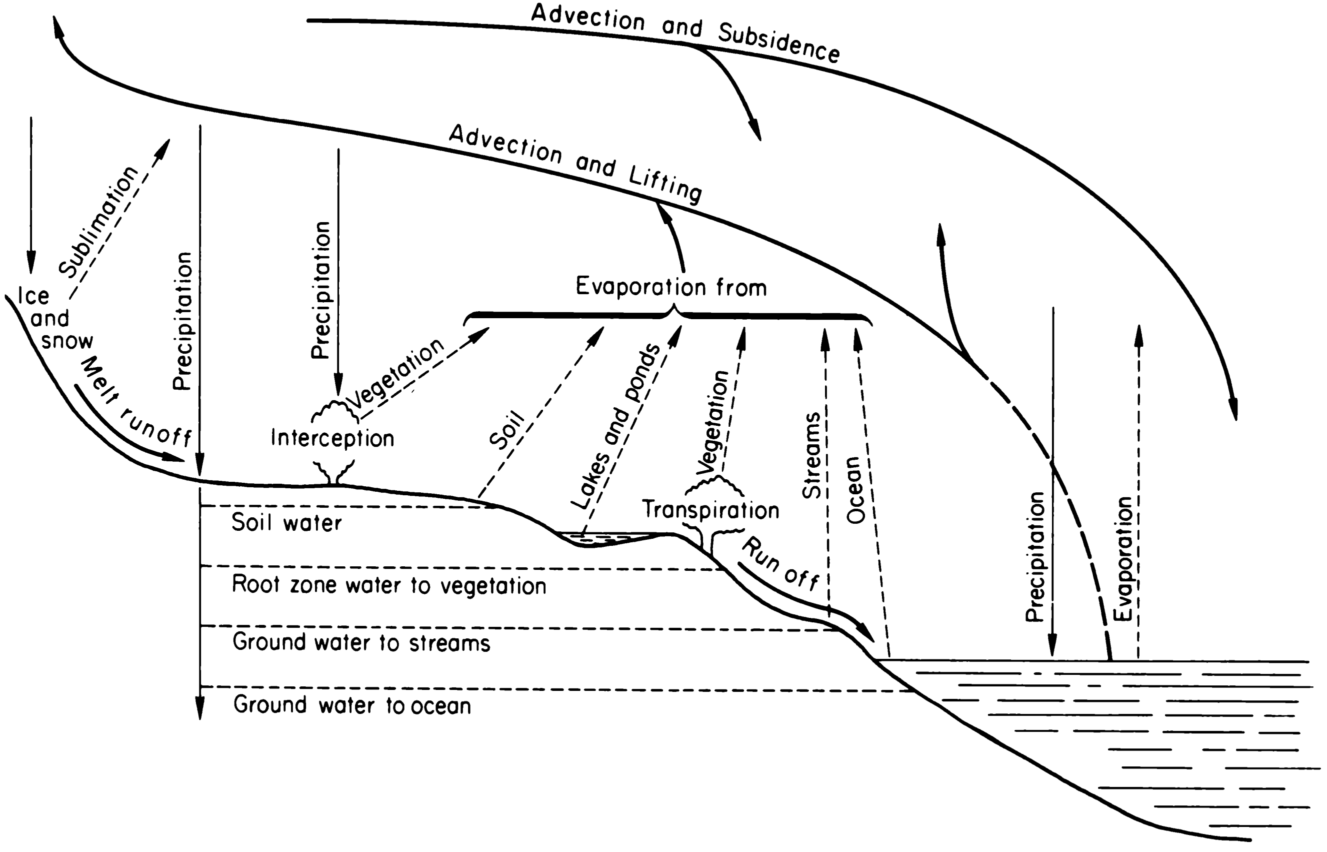


Figure 11.1
The hydrologic cycle.

precipitation in polar regions. Insolation, through its effects on the atmospheric processes, provides the "fuel" to operate the hydrologic cycle. The flow of water vapor across boundaries of watersheds or continents is just as much a part of the cycle as the flow of a river to the sea.

The practical value of an understanding of the hydrologic cycle comes from the fact that it is the means by which water is made available on the land for the many uses by mankind and other life forms. A great deal of mental and physical effort is expended in trying to control the cycle to our better advantage. The conservation of water resources, indeed of most natural resources, is intimately related to one or more of its phases. Where we cannot control it we must adjust to its peregrinations. Rainmaking cannot increase the total water supply, but it might speed up the hydrologic cycle so that at a specific time and place more water is available for immediate use. In its broadest sense, water conservation aims to slow down and use more effectively the runoff phase of the water cycle and to accelerate and redistribute the atmospheric phase.

THE WATER BUDGET AT THE EARTH'S SURFACE

The major demands for water arise at the earth-atmosphere boundary, which is also the zone of maximum energy and moisture exchange between the surface and the air. Accordingly, we can consider relevant elements of the hydrologic cycle by means of an accounting procedure—the water budget of the earth's surface. The essential elements of moisture exchange for a given land area and specified period of time are combined in the equation:

$$P = ET + dST + S$$

in which P is the income from precipitation, ET is the loss by evapotranspiration, dST is the gain or loss of storage in the soil, and S is surplus. Thus, all of the income is accounted for by expenditures, an increase or decrease in savings, and a possible surplus that requires wise management. The transfer of moisture to the air by evapotranspiration is a function of both available water and available energy. The amount that could be evaporated and transpired by the available energy is the potential evapotranspiration (PE), or simply the water need. (See pages 148–51.) The amount lost upward from the surface is the actual evapotranspiration (AE). It is equivalent to ET in the water budget equation and therefore is limited by precipitation and soil moisture storage. When total precipitation exceeds potential evapotranspiration part of the water may be needed to restore soil moisture; the remainder, if any, is surplus that may run off, percolate to the ground-water level, or accumulate in ponds. If *PE* exceeds *AE* and soil moisture is being depleted there is a deficit (*D*), and the equation becomes:

$$P = PE - dST - D$$

Expenditures fail to meet the need imposed by the energy-moisture system. Obviously the income is inadequate, although savings may help to offset a temporary shortage. Table 11.1 shows an example of water budget accounting on a monthly basis for a representative station in a humid subtropical climate. The same data are displayed graphically in Figure 6.1.

TABLE 11.1
Monthly Water Budget — Memphis, Tennessee (all values in millimeters)

	J	*F*	*M*	*A*	*M*	*J*	*J*	*A*	*S*	*O*	*N*	*D*	*Yr*
P	134	112	131	126	106	94	84	80	72	72	107	114	1,232
PE	5	10	31	65	113	152	173	160	114	64	26	10	923
*ST**	300	300	300	300	293	241	179	137	119	127	208	300	
dST	0	0	0	0	-7	-52	-62	-42	-18	+8	+81	+92	
AE	5	10	31	65	113	146	146	122	90	64	26	10	828
D	0	0	0	0	0	6	27	38	24	0	0	0	95
S	129	102	100	61	0	0	0	0	0	0	0	12	404

*Amount of soil moisture storage at end of each month, assuming a capacity of 300 mm.

The water budget concept can be applied in the analysis of the moisture exchange at any surface—a plant leaf, a forest canopy, a glacier, an entire continent, or an ocean. It is an effective means of determining drought conditions and the proper use of irrigation water. One of the most fruitful applications is in the predictions of runoff (surplus) from river basins under varying climatic conditions. Similarly, ground-water supplies (another form of surplus) may be calculated. Any single element in the budget can be determined mathematically if all the other elements are known.

EVAPOTRANSPIRATION

The accuracy of water budget accounting depends in turn on the accuracy of the respective values used in the computation. Precipitation and its measurement have been discussed in Chapter 3. It is important to note that precipitation gauges provide reliable totals for large areas (e.g., entire river basins) only if placed at representative sampling sites.

The key element in the water budget is evapotranspiration, which is also the link between moisure and energy exchanges. Attempts to measure evapotranspiration directly involve humidity gradients and vapor transport by diffusion or turbulent motion. Because humidity decreases with distance from the evaporating surface, the rate and amount of evapotranspiration can be determined if the difference in water vapor content and the rate of vertical mixing through the surface layer of the air are known. Special instruments and plant chambers assist in obtaining the necessary data.

An indirect method of measuring evapotranspiration depends on the water budget concept itself and employs the drainage lysimeter, or *evapotranspirometer* (Figure 11.2). The essential features of this device are (1) a sunken field tank filled with soil in which plants like those of the surrounding area are grown and (2) a covered percolation tank sunk into the ground a short distance from the field tank to catch surplus water that drains from the field tank through (3) a connecting underground tube. The field tank receives water only from precipitation or irrigation and loses it by downward percolation or by evapotranspiration. For determination of potential evapotranspiration the field tank and the surrounding area are irrigated whenever necessary to maintain optimum soil moisture conditions. When the amounts of water added by rain and auxiliary watering are known, potential evapotranspiration can be found by subtracting the water loss by percolation from the total received in the evapotranspirometer field tank. Soil moisture storage may vary

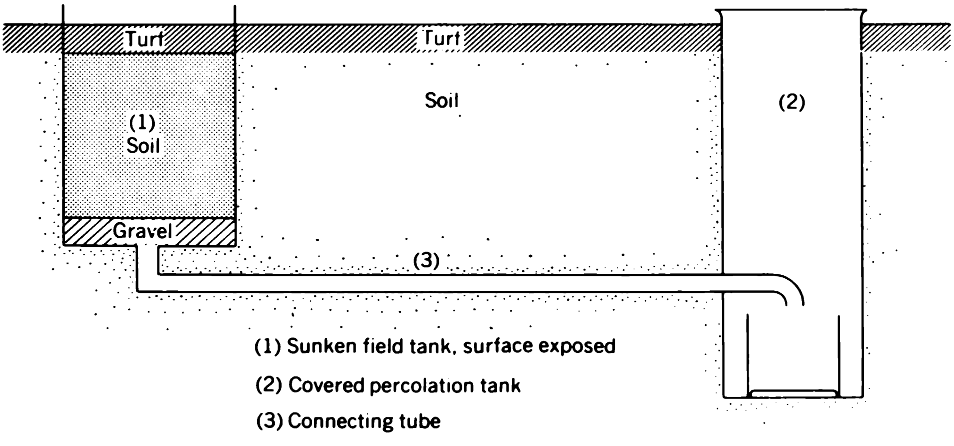

Turf
Turf
(1)
Soil
Soil
(2)
Gravel
(3)
(1) Sunken field tank, surface exposed
(2) Covered percolation tank
(3) Connecting tube

Figure 11.2
Essentials of a field evapotranspirometer. (Adapted from John R. Mather, *The Measurement of Potential Evapotranspiration*, Johns Hopkins Laboratory of Climatology, 1954)

from day to day, and the amount of water collected in the percolation tank varies with it, but over longer periods the fluctuations are not critical.

Expensive but more accurate versions of the evapotranspirometer are weighing lysimeters that provide a constant record of changing water content in the soil and measurements of actual evapotranspiration for periods of only a few minutes. As in the case of precipitation gauges, problems arise in selecting and maintaining representative sites for evapotranspirometers. Standardization of plant cover to provide comparability of transpiration values is also difficult. Artificial watering of sites, especially in dry areas, in order to achieve maximum (potential) evapotranspiration creates an "oasis effect," which might be quite different from the prevailing regional climate. Different slopes and exposures introduce further problems in the interpretation of data.

Because of the rather complex equipment required for measurement of potential evapotranspiration and the attendant difficulties in assuring representative conditions, estimates are often based on the records of actual evaporation from open pans or other types of evaporimeters. Mathematical relationships among moisture, energy, and air movement factors also yield useful approximations to water need. The method developed by Thornthwaite uses temperature and the length of the daylight period as energy expressions. (The *PE* values in Table 11.1 were calculated using his formula.) H. L. Penman in the United Kingdom and M. I. Budyko in the Soviet Union have refined the energy budget approach to potential evapotranspiration by incorporating net radiation as well as factors for wind speed and humidity. Although details of these and other theoretical methods are beyond the scope of this book, they offer greatly improved understanding of the physical bases of climate.

SOIL MOISTURE AND GROUND WATER

The role of soil moisture in the water budget is highly variable; it depends not only on the other budget factors but also on the capacity of a soil to hold water. The amount of water a saturated soil can retain against the pull of gravity is its *field capacity*. It varies mainly with soil texture but also with structure, organic matter content, and the depth of the solum. Fine clays have high field capacities, whereas sandy soils hold little moisture. For agricultural purposes it is useful to consider the minimum amount of soil water that is necessary in the root zone to allow extraction by plants. This minimum is the *wilting point,* and it depends on the same factors that govern field capacity. Relative field

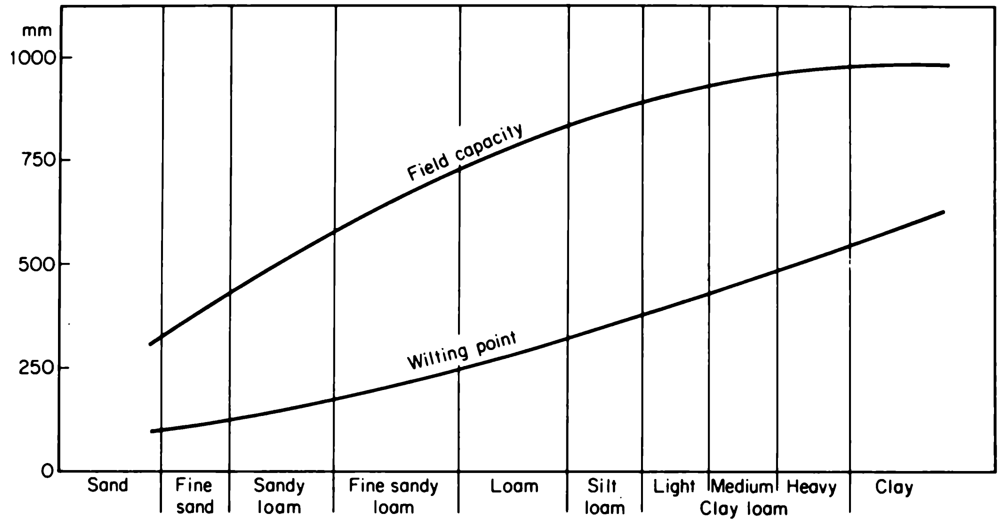


Figure 11.3
Average field capacity and wilting point for different soil textures. Values are in millimeters per 300 cm depth of soil in the plant root zone. (After Smith and Ruhe, *Water, 1955 Yearbook of Agriculture,* p. 120)

capacities and wilting points for various soil textures are shown in Figure 11.3.

The water content of soil can be ascertained by several methods, none of which is completely satisfactory for all purposes. A simple but time-consuming procedure is to weigh soil samples before and after drying in an oven. Unfortunately the samples cannot be reused. Weighing lysimeters accurately measure changes of soil moisture in a specific volume of soil. Another device, the *soil moisture tensiometer,* is a porous clay cell filled with water and placed in the soil. A pressure gauge attached to the cell registers changing tension as water moves through the cell toward drier soil or into the cell from wet soil. An electrical method depends on the relationship between soil moisture and the conduction of electricity by buried resistance blocks made of gypsum, nylon, or fiber glass. It is especially useful for remote recording. A neutron scattering method relies on a slowing of fast neutrons by collision with hydrogen atom nuclei. A probe containing a source of fast neutrons and a detector of slow neutrons is lowered into the soil. Since most of the hydrogen is in soil water molecules the number of slow neutrons detected is an index of soil moisture.

When the soil zone has reached field capacity excess soil water percolates downward under the force of gravity through a subsurface zone of aeration (vadose water) to a zone of saturation known as *ground water.* Depths to the upper surface of ground water, that is, the water table, can be measured in wells and are used to estimate ground-water supplies. Ground water may also be treated as part of the surplus in the water budget, and when other values in the budget (including runoff) are known the amount transported to ground water can be calculated.

RUNOFF AND FLOODS

Runoff is that portion of precipitation that returns to the oceans and other water bodies over the land surface or through the soil and water table. It may involve the direct return of rainfall or the flow of melted snow and ice fields which have temporarily stored the water. The significance of runoff to soil erosion has been discussed in Chapter 10. Floods differ from simple runoff only in degree, the distinction between the two depending for the most part upon how they affect surface features. River floods result whenever the channel capacity is exceeded by the runoff. Excessive runoff of rainfall or snowmelt is the fundamental cause, but the channel capacity may also be affected by barriers to flow (such as dams or ice jams), sudden changes of direction of the stream, reduced gradient, siltation of the stream bed, or sudden release of water due to a broken dam. On many rivers, floods are defined in relation to arbitrary gauges placed at important points along the stream. Not all floods are "bad." Many of the features of the earth's surface have been sculptured by running water. Since antiquity agricultural areas in the lower-Nile flood plain and Mesopotamia have depended upon annual river flooding and the accompanying deposits of fertile silt. What is gained in this way in the lowlands must be lost at higher levels in the watershed.

The amount of runoff in a given region is conditioned by several factors: the amount and intensity of precipitation, temperature, character of the soil, vegetative cover, and slope. If precipitation occurs as rain, the proportion that runs off will depend upon the capacity of soil and vegetation to absorb it. Plants have the ability to retain some rainfall on their external structures and to slow the velocity of raindrops. They also detain water in its horizontal movement. They improve soil structure and their roots provide channels to divert excess soil water into the ground water at greater depths. The high humus content of soils with a dense grass cover enhances absorption, for it acts something like a sponge. Porous soils absorb more water by infiltration than dense clays or rock. If already saturated or if sealed by the intense pounding of raindrops, however, their absorptive capacity is limited. Similarly, impervious subsoil reduces the amount of water that can be taken in. When frozen in a wet condition, soils can absorb very little rain, but if frozen when porous and nearly dry they have a high capacity for water.

Surface runoff varies with the character of an individual storm.

Long-continued rains charge the soil to its full capacity and the proportion which runs off becomes progressively greater. Loose sands and gravels constitute an exception, if they are not underlain with impervious material. In the case of snow or hail, infiltration must await melting. When the soil and plant cover are warm, some melting begins at once. If the surface and the air above are too cold, there is a delay of hours, days, or even months in the runoff. Ultimate surface flow is determined by the various surface and soil characteristics, the rate of melting, and the amount of snow lost by sublimation. Sometimes snow and ice are removed principally by sublimation, or by melting and evaporation, so that the runoff phase of the water cycle is assumed partly by the air. This removal process is referred to as *ablation.*

CLIMATIC CAUSES OF FLOODS

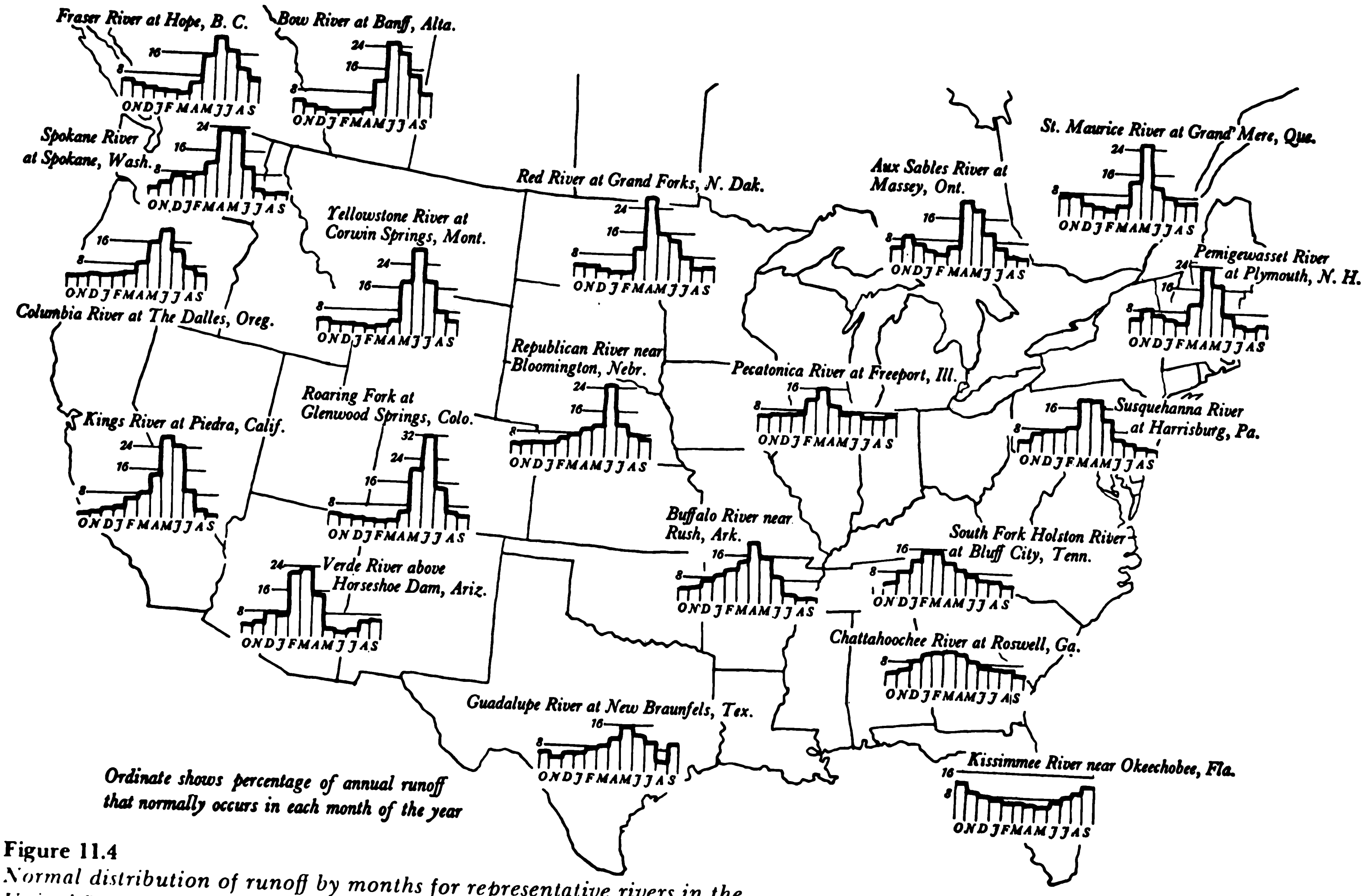
The predisposition of a climate to storms producing excessive precipitation is the fundamental basis of the flood hazard. In some climates flood-producing storms occur irregularly; in others they follow a seasonal pattern. Two types of storm are the initiating agents for most rain-caused floods: the violent thundershower, which is of short duration and produces flash floods, and the prolonged, general rain which, through sheer quantity of water, creates extensive flooding over entire watersheds. The flash flood is common only in those regions which experience heavy thunderstorms, that is, where unstable maritime air masses are frequent visitors. However, the occasional thunderstorms of arid climates also lead to destructive floods. The thunderstorm may be of convective, frontal, or orographic origin; so far as flooding is concerned, the important feature is its intensity, not its origin. In general the less a flash flood is expected, the more disastrous it is likely to be. In the desert it is a decided anomaly; in the rainy tropics it is anomalous to the extent that the normal stream channels may be unable to carry away the excess runoff. Although scattered convective storms or a line of frontal thunderstorms often contribute to extensive flooding in large river systems, such floods are usually the result of long-continued, warm, rainy weather. This type of weather is most common in the humid climates of middle latitudes when frontal systems occlude or remain stationary for protracted periods, but it is accentuated dramatically in hurricanes. As long as the storm system is fed by moist maritime air, rain can continue. Once the water-holding capacity of soil and vegetation is exceeded, the foundation is laid for a major flood. Such conditions are typical of the Ohio Valley in spring, when neither the wet soil, the swollen tributaries, nor the high ground-water table can absorb even a moderate fall of rain.

If the precipitation falls as snow, the flood is not necessarily avoided. It is merely delayed. The most effective instrument for snow removal is a warm wind. If it is dry (for example, the foehn wind), it carries much of the water away in the air and is not a likely cause of flooding; if it is moist as well as warm, it produces rapid melting and runoff. Direct sunshine is not a major factor in rapid melting owing to the high albedo of snow. Rainfall on snow does not at once produce runoff of flood intensity, for a snow cover has considerable capacity to hold water. When prolonged rains come in combination with warm winds, however, ideal flood conditions prevail. Another factor in snowmelt runoff is the temperature of the soil under snow cover. If the soil is frozen there is very little melting and runoff under the snow, but melting due to other causes is likely to produce a high proportion of runoff. Unfrozen soil causes some melting throughout the winter, and a large part of the water enters the soil to become ground water rather than runoff, because the melting process is slow.

The seasonal distribution of precipitation is the principal determining factor in the regimes of rain-caused floods. In the arid climates the thundershowers and resulting flash floods are both erratic. Concentrations of rainfall in one season (for example, as in the monsoon tropics) are likewise accompanied by seasonal floods. In mountainous areas of mid-latitudes and at high latitudes, flooding due to snowmelt runoff occurs with the onset of the warm season and is not always directly related to the precipitation regime. For example, in the dry summer subtropics the precipitation maximum is in winter, but the maximum runoff of melted snow is in spring or early summer. On the other hand, spring and summer snows in the mountains of marine climates contribute to the runoff in those seasons, often being removed by subsequent rains. In humid continental climates seasonal floods come at the end of the period of soil moisture accumulation, that is, in spring.

Destructive floods occur in every month in the United States, often with loss of life and damage to property running into many millions of dollars. In the United States, as a whole, loss of property and deaths due to floods have been great in all months, with winter and spring the worst seasons. The normal maximum of runoff of streams shows a distinct seasonal pattern. Figure 11.4 shows the normal annual distribution of runoff for representative rivers in the United States and southern Canada. Note the influence of late snowmelt on the regimes of northern rivers. In Florida snow is not a factor in runoff and the runoff is greatest in autumn in response to the rainier season and decreasing evaporation.

A variation in the normal runoff regime of snow-fed rivers occurs in the Mackenzie, Ob, Yenesei, and other rivers that flow northward at high latitudes in North America and Asia. Melting begins in the headwaters and middle courses of these streams earlier in the spring than it


Fraser River at Hope, B. C.
Bow River at Banff, Alta.
Spokane River at Spokane, Wash.
Columbia River at The Dalles, Oreg.
Yellowstone River at Corwin Springs, Mont.
Kings River at Piedra, Calif.
Roaring Fork at Glenwood Springs, Colo.
Verde River above Horseshoe Dam, Ariz.
Red River at Grand Forks, N. Dak.
Republican River near Bloomington, Nebr.
Guadalupe River at New Braunfels, Tex.
Buffalo River near Rush, Ark.
Pecatonica River at Freeport, Ill.
Aux Sables River at Massey, Ont.
St. Maurice River at Grand' Mere, Que.
Pemigewasset River at Plymouth, N. H.
Susquehanna River at Harrisburg, Pa.
South Fork Holston River at Bluff City, Tenn.
Chattahoochee River at Roswell, Ga.
Kissimmee River near Okeechobee, Fla.
O N D J F M A M J J A S
8
16
24
32

Ordinate shows percentage of annual runoff that normally occurs in each month of the year

Figure 11.4
Normal distribution of runoff by months for representative rivers in the United States and Canada. (From *Water, 1955 Yearbook of Agriculture*)

does in the lower flood plains. Because the downstream channels are choked with ice there is extensive flooding until the thawing season is well underway. A contributory factor is permafrost, which allows little or no infiltration of surface waters.

In summary, floods are a temporary acceleration of the runoff phase of the hydrologic cycle. They can always be traced back to moisture in the atmosphere, but their intensity is affected by both natural and cultural features of the land surface.

RUNOFF FORECASTING

Runoff forecasting has two principal phases: One is concerned with forecasting the amount of rainfall or snowmelt that will be available to run off; the other attempts to predict the rate of flow in relation to stream capacity. Figure 11.5 summarizes graphically the main factors which are considered in flood forecasting for a particular storm. The techniques of forecasting quantitatively the precipitation to be expected from a storm are being refined, but, generally speaking, only qualitative indications of impending distaster due to heavy rainfall can be given in advance of the storm. For example, if a cyclone is drawing large quantities of tropical maritime air over a watershed which is already wet there is a distinct possibility of flooding. Extreme instability of the

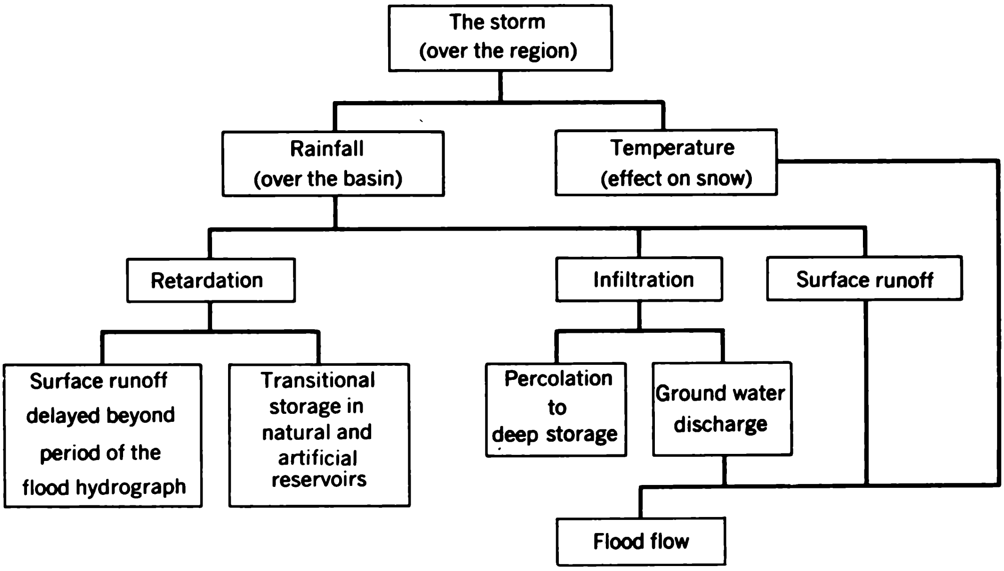


Figure 11.5
Factors involved in runoff forecasting. (From *Climate and Man, 1941 Yearbook of Agriculture*)

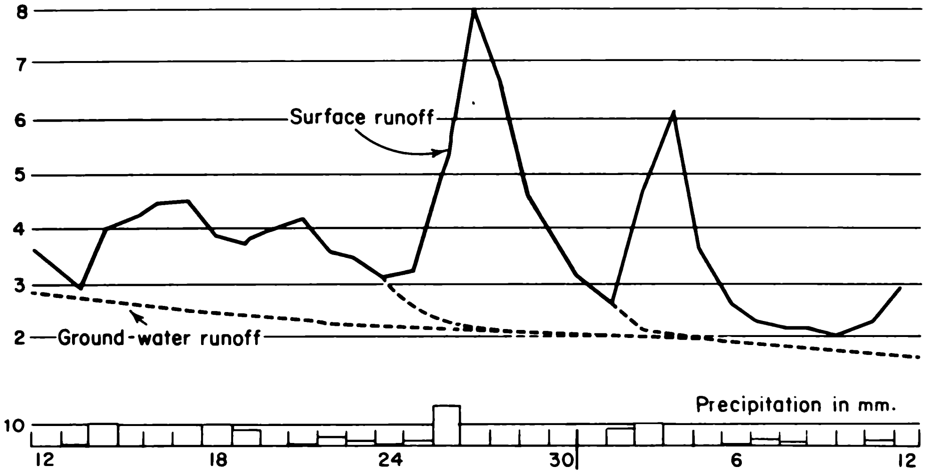


Figure 11.6
Hydrograph of stream flow in the Muskingum River at Dresden, Ohio, during a 31-day period. Values on the left represent relative discharge volume. The lower graph shows daily precipitation in millimeters. (Adapted from *Water, 1955 Yearbook of Agriculture*)

moist air may presage an even greater flood threat. Temperatures of the conflicting air masses within the storm, the temperature of air expected to follow the storm, and wind are factors which modify the meteorological aspects of the flood hazard. If the precipitation is snow, the threat is deferred until melting begins. Forecasts of snowfall are not of such immediate importance in flood warnings as are forecasts of rainfall and potential snowmelt. When there is some question of whether rain or snow will fall generally over a watershed, the flood forecaster is in a delicate position. In mountainous areas, the lower level of snowfall will determine the area to be affected by rainfall runoff. Thus, a careful investigation of the probable upper-air temperatures is necessary. In regions visited by hurricanes, advance warning of the coming storm is a prerequisite to timely forecasts of so-called "storm surges" along coasts as well as of excessive runoff inland.

Measurements of rain even as it falls provide valuable data for flood prediction. Networks of precipitation-gauge stations have been set up in many watersheds of the United States. Correlated with subsequent stream flow, these measurements help to establish patterns of rainfall-flood relationships that are the basis for future flood forecasts as well as of immediate utility. *Hydrographs,* that is, graphs on which stream flow is plotted as a function of time, have been useful devices for studying the effects of storms on flooding. (See Figure 11.6.) The progress of storm centers across the river basins, the characteristics of air masses, the measured rainfall over the watersheds, and past stream behavior are the principal factors employed in making stream-flow forecasts.

After precipitated waters begin to flow from a watershed, much more accurate forecasts of stream discharge are possible. Such forecasts are based primarily on a knowledge of the flow patterns of main streams and their tributaries. Reports of measured flow from upstream locations are collected to determine the amount of water that can be expected to pass specific points downstream. One swollen tributary does not produce a major river flood, but several together may. In the headwaters, the period of advance warning is shortest and reliance on rainfall forecasts is greatest. Downstream forecasts are more accurate and there are usually several hours advance warning. Exceptions occur when torrential rains fall in lowlands or when major obstructions to runoff, such as flood-control dams, break.

SNOW SURVEYING

In the mountainous regions of the world, especially in middle latitudes, nature provides a reservoir for storage of water in the form of snow to be released with the rising temperatures of spring. This fact is of vital significance to water users in arid lands adjacent to snow-capped mountains. Efficient planning for water use (and control) during the season of peak flow is better carried out with a prior knowledge of expected stream flow. As yet it is not possible to predict accurately far in advance how much snow will fall on mountain snowfields. The alternative is to measure the water content of the snowpack so that potential runoff can be determined.

Each winter several public and private organizations in the western United States cooperate in snow surveys with the Soil Conservation Service as the coordinating agency. The principle of snow surveying involves a sampling technique, since quantitative measurements of all the snow at upper levels is neither feasible nor necessary. Representative areas known as *snow courses* are selected in the snowfields which feed the major watersheds. It may be necessary to select several snow courses for a region to cover the differences in storm frequency and intensity which occur. Measurements on the snow courses are taken on the first of each month throughout the winter and spring. Runoff forecasts are prepared early in the winter and revised monthly. The depth of snow alone is inadequate for snow surveys used in runoff forecasting. The water content is the important factor, for snow densities vary greatly. A snow sampling device permits measurement of water content as well as of depth. Coupled sections of lightweight aluminum tubing are thrust

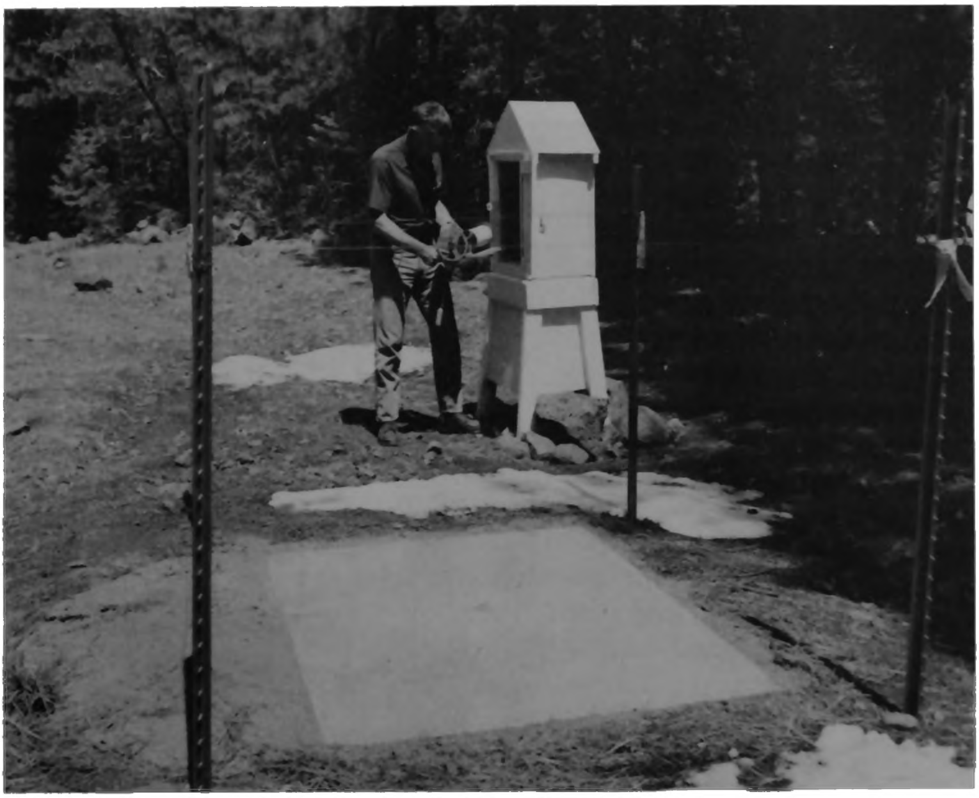


Figure 11.7
Snow pillow installation and shelter for recording instruments. (Soil Conservation Service—USDA)

vertically downward through the snow to the ground beneath. A serrated cutting edge facilitates penetration through ice or hard snow layers. Snow weight, that is, water content, is determined by weighing the snow and tube on a simple spring scale, which is calibrated to make allowance for the weight of the tube. Snow surveyors take 10 or 12 samplings at intervals of 15 to 30 m along each snow course. Personnel from several cooperating agencies reach the snow courses on skis, by snow tractor, or in light planes or helicopters.

The *snow pillow* is another technique for sampling snow water content. Accumulated snow on a flat, liquid-filled container (pillow) creates internal pressure and activates an index calibrated in terms of snow water depth (see Figure 11.7). Measurements may be recorded electrically or transmitted as radio signals from remote locations. Some permanent remote-control stations incorporate the principle of radar transmission between a transceiver mounted on a pole and a target at ground level.

A method that permits sampling at different sites uses radioactive isotopes. The gamma rays from the cobalt-60 isotope, for example, are transmitted through snow in proportion to the water content of the snow. When the isotope is inserted to the bottom of a snow layer, measurement of the gamma-ray intensity at the surface yields an index of water content that can be converted into depth of water.

Through the correlation of the snow survey information with subsequent runoff from the respective watersheds, formulas and graphs to express the relationship have been established. These are fundamental to water supply forecasts which, as a result, have become increasingly accurate. Such forecasts give a good indication of the amount of snowmelt which will enter the streams of a given watershed, but they cannot tell when the runoff will occur or whether there will be destructive floods. When full-scale runoff begins, because of high temperatures or storms, the flood forecaster must resort to the techniques described in the previous section, but information on the amount of water in the snowpack is valuable for assessing the imminence of a major flood. Besides aiding flood forecasting, the snow surveys provide information on the probable water supply for irrigation, hydroelectric power, navigation, fisheries, and a host of other water uses. An irrigation farmer, knowing in advance what the water supply will be, can plan his crops accordingly; power companies can compute future output and make firm commitments to consumers; and the industrial consumers of electric power can expand or curtail their production with a minimum of financial loss. City water departments are able to establish realistic sprinkling regulations; the fish and wildlife agencies can plan with greater accuracy for winter feeding and summer water supplies for fish and waterfowl on the basis of snow-cover information. Snow survey data have been used, with some success, to predict the ground-water supplies of certain river basins.

Special surveys in recreation areas gather supplementary data on conditions for skiing and avalanche forecasts. Unstable snowpacks on steep slopes can be a hazard to transportation and communication facilities, winter sports enthusiasts, and buildings. The danger is so acute each winter in Switzerland that a special Swiss Avalanche Commission has been organized. Avalanches are, in a sense, a flash flood of water in the solid form (see Figure 11.8). While the angle of repose of the snow cover and the underlying surface features are basic considerations in avalanches, they are set off naturally by unfavorable conditions of snow structure, temperature, humidity, and wind. Any disturbance of an unstable snow cover is likely to cause an avalanche. *Moist* snow is relatively cohesive and does not avalanche readily, but snow *soaked* with water is predisposed to flow or slip. New snow, composed of light, interlocking snow crystals, is stable, as is granular snow, whose particles resist movement. The layering of the snowpack also influences the tendency to avalanche. Generally, the layers reflect weather fluctuations. Some layers

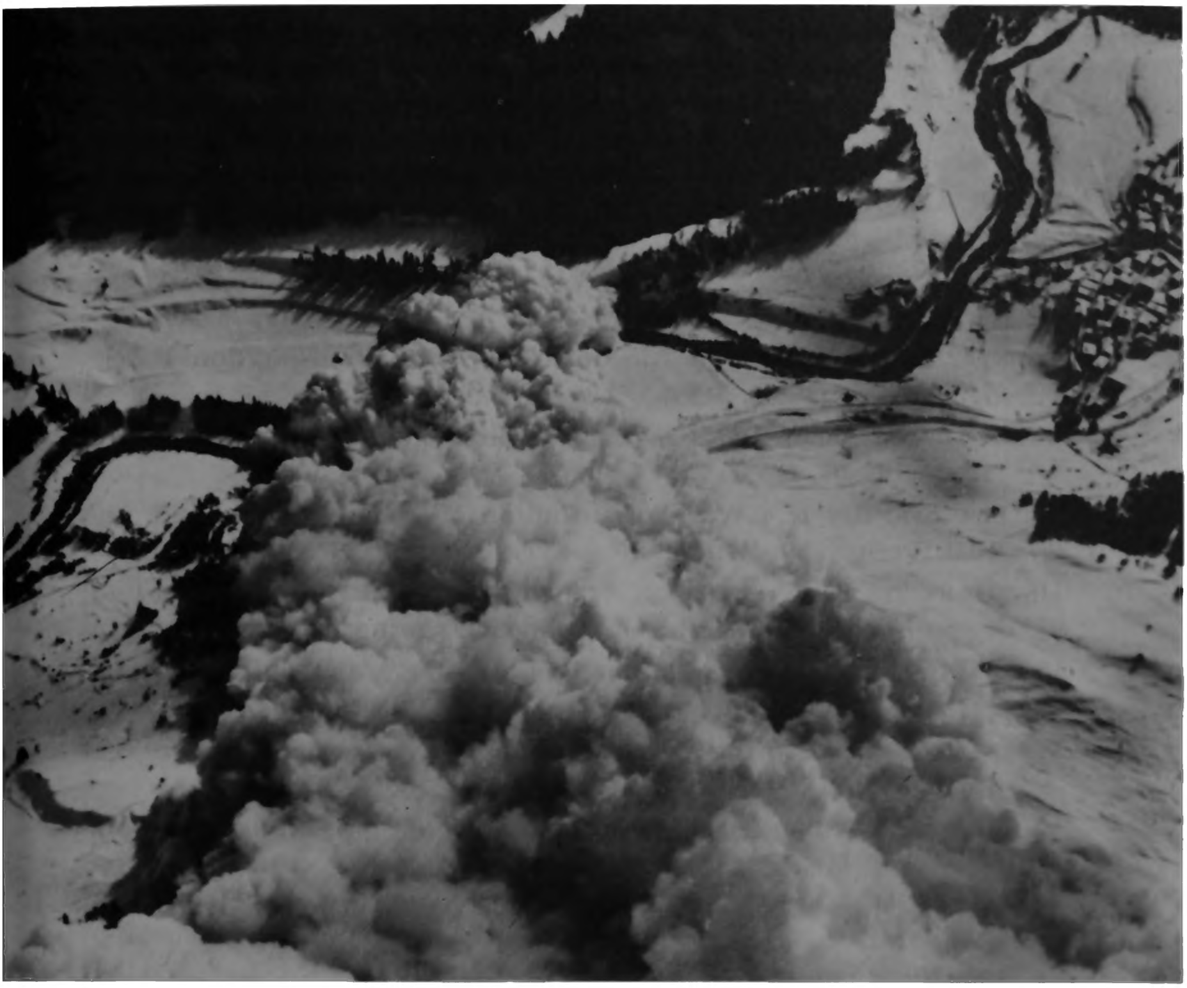


Figure 11.8
Snow avalanche in a Swiss valley. (Courtesy Institut fédéral pour l'étude de la neige et des avalanches, Suisse)

may be wet and compact; others are loose and dry. Snow cores provide data on the snowpack strata; meters measure the pressure of the snow, and rams test its resistance and coherence. Avalanche hazard forecasts take into account the amount and characteristics of snow, terrain factors, and the current and predicted weather conditions.

Wind blasts known as *avalanche winds* frequently accompany severe avalanches. They are generated by the mass movement of snow against the air and by greatly reduced pressure behind the moving snow. The more violent blasts have been known to break off trees several meters away from the main path of the avalanche.

Additional Readings

CRITCHFIELD, HOWARD J., "Water Balance Analogues in the Marine Climates of New Zealand and North America," *N. Z. Geogr.,* 22, 2 (1966), 111–24.

FOX, R. C., "An Oil-Wax-Surfactant System for Retarding the Evaporation of Water," *Nature,* 205 (1965), 1004–5.

LAMER, VICTOR K., AND THOMAS W. HEALY, "Evaporation of Water: Its Retardation by Monolayers," *Science,* 148, 3666 (1965), 36–41.

SEGINER, I., "The Effect of Albedo on the Evapotranspiration Rate," *Agri. Met.,* 6 (1969), 5–31.

THOMPSON, HERBERT J., "The Black Hills Flood," *Weatherwise,* 25, 4 (1972), 162–67, 173.

TULLER, STANTON E., "World Distribution of Mean Monthly and Annual Precipitable Water," *Mon. Wea. Rev.,* 96, 11 (1968), 785–97.

UNITED STATES DEPARTMENT OF AGRICULTURE, *Water, 1955 Yearbook of Agriculture.* Washington: Government Printing Office, 1955.

VESTAL, C. K., "The Precipitation Day Statistic," *Mon. Wea. Rev.,* 89, 2 (1961), 31–37.

WARD, R. C., "Measuring Evapotranspiration; A Review," *Journ. Hydrol.,* 13 (1971), 1–21.

WEBB, E. K., "A Pan-Lake Evaporation Relationship," *Journ. Hydrol.,* 4 (1966), 1–11.

WIESNER, C. J., *Hydrometeorology.* London: Chapman and Hall Ltd., 1970.

World Water Balance, Vols. 1–3, Proceedings of the Reading Symposium, July, 1970. Gentbrugge, Belgium: International Association of Scientific Hydrology, 1972.

chapter twelve

Climate and Agriculture

As climate largely defines the climax vegetation formations throughout the world, so it sets limits for crop production. Domestic animals, too, respond to climatic differences, both physiologically and through their feed requirements for economic production of food and fiber. In general, crops and animals have their optimum climatic conditions for production, although we must not lose sight of other variables such as soils, relief, insects and diseases, market, and transportation facilities which interact to modify the suitability of a particular area for a specific type of cultivation or animal husbandry. The actual distribution of crop plants is determined by the combined influence of physiological, economic, social, technological, and historic forces; but no crop can attain importance in an agricultural system unless it is adapted to prevailing environmental conditions.

General aspects of the relationship of climate to crops have been known for centuries; yet important details of the influence of climate on agriculture are still being revealed. In examining the relationship, we are interested in two broad, overlapping problems. One is concerned with the effect of climate on the distribution of crops and domestic animals; the other pertains to the more detailed influence of climatic elements on specific crops or animals and their productivity. Agriculture is one of the riskiest of all enterprises. Not the least of the risks are those imposed by weather and climate.

CLIMATIC FACTORS IN CROP PRODUCTION

The principal climatic factors affecting crop production are the same as those influencing all vegetation—temperature, length of growing season, moisture conditions, sunlight, and wind—but they must be considered in a different light with respect to crops. Natural vegetation is, axiomatically, adapted to the climatic conditions with which it is associated. For economic reasons, man has domesticated crop plants and cultivated them in environments where they could not survive without his help. In other words, crop plants are less hardy than natural vegetation. Agriculture is essentially a combination of processes designed to promote, artificially, a favorable environment for growth. Its practices are so diverse as to include drainage and irrigation to control the moisture factor, artificial sheltering to control temperature, and shading or the use of electric lights to control light. In the extreme, any crop can be grown anywhere if labor and expense are not in question; in the realm of practicability, all crops have their climatic limits for economic production—limits which can be extended by plant breeding and selection as well as by cultivation methods.

In their influence on crops, the climatic factors are closely interrelated. The effect of each is modified by the others. Daily, seasonal, or annual variations in any or all of the climatic elements are of importance in determining the efficiency of crop growth. In considering the climatic environment of plants it is important to keep in mind that the microclimate immediately around the plant is of vital significance. Critical conditions, whether they be at the maximum or minimum limits for the plant, may prevail in the air at plant level while at a short distance above the surface the temperature is well within the safe range. Similarly, moisture, light, and wind effects may be quite different at and near ground level from conditions in the free air above.

TEMPERATURE AND CROPS

The temperature of the air and of the soil affects all the growth processes of plants. Every variety of every crop plant has minimum, optimum, and maximum temperature limits for each of its stages of growth. Winter rye has relatively low temperature demands and can withstand freezing temperatures during a winter period of dormancy. Tropical crops (for example, dates or cacao beans) have high temperature requirements throughout the year. The upper lethal temperatures for active plant cells of most species range from 50° to 60°C but vary with species, stage of growth, and length of exposure to the high temperature. High temperatures are not as serious as low temperatures in arresting plant development—if the moisture supply is adequate and the crop is adapted to the climatic region. Crops ordinarily do not "burn up"; they "dry up." But, under very high temperatures, growth is slowed or even stopped regardless of the moisture supply, and premature loss of leaves or fruit is likely. Mid-latitude plants will *grow* structurally but *fruit* poorly in the tropics. Disaster to crops usually comes with the combination of dry and hot conditions. Winds that would be expected to produce evaporative cooling often only speed up transpiration and result in dehydration of plant tissues. The optimum temperature for the maximum rate of plant growth is not always the best for crop production. Temperatures that promote quick growth may also result in weak plants which are more easily damaged by wind, hail, insects, or disease. Some crops fail to produce their fruit or seed if forced under high temperatures. Within limits, the problem of high temperatures can be solved under field conditions by increasing the moisture supply through irrigation or moisture-conserving tillage practices. These practices permit the successful production of many different crops in hot desert oases. Delicate plants can be shaded from the direct rays of the sun by higher tree crops or artificial cloth or slat shades. Selection of planting sites on the less-exposed slopes of hills can also alleviate high temperatures if the crop demands it. Sometimes it is desirable to have low temperatures during the stages of germination or flowering of crops. This is achieved by planting early in the season or in a cooler location, or, as in the case of winter wheat, by planting before the cold season. Water bodies defer the rise of temperatures to the leeward in spring and thereby help to prevent blossoming in fruit orchards until the danger of killing frost is past. Such is the case in the fruit-producing areas east of the Great Lakes. Citrus fruits take on better coloring and become sweeter when subjected to short periods of near-freezing temperature. Strawberries develop their best flavor when daytime temperatures are about 10°C.

Few crops are restricted in their worldwide distribution because of the direct effects of high temperatures alone, although many yield best when grown near their poleward limits. Since production of many crops has been pushed farther and farther into high latitudes, the effects of low temperature have had increasing significance. In discussing the effects of low temperatures on crops, we must distinguish between chilling and freezing. Prolonged chilling of plants at temperatures above freezing retards growth and can kill certain plants which are adapted only to constantly warm conditions. Chilling may not directly kill plant cells but reduces the vital flow of water from the roots so that transpiration losses cannot be regained. Yellowing of plant leaves sometimes results from this kind of physiological drought. Crops such as rice and cotton are killed by near-freezing temperatures of two or three days' duration; potatoes, maize, and many garden vegetables can "weather" such cold spells with little or no damage, although their growth is slowed. As would be expected, the warm-climate crops are most seriously affected by chilling. Application of cold irrigation water to a field retards growth by reducing temperatures in the soil and at the immediate surface. Evaporation of soil moisture also tends to decrease the temperature. Loss of heat due to evaporation may be serious in the case of flooded rice paddies. In northern Japan chemical solutions which act as evaporation suppressors are applied to the surface of flood waters in order to maintain higher daytime temperatures at the level of the growing rice plants. On the other hand, wet soils do not cool as rapidly at night, and irrigation water can actually help prevent chilling if its temperature is above that of the air.

The influence of low night temperatures varies greatly with different crops. Certain crops, such as potatoes and sugar beets, store carbohydrates more rapidly during periods with cool nights. Cotton, maize, and tobacco require warm nights for maximum development. In high latitudes, the greater length of the daylight period makes it possible for crops otherwise adapted to cool climates to condense their growing season into fewer days. Light as well as heat contributes to speeding growth during the long days.

Whenever growing plants are subjected to freezing temperatures, damage or even death is imminent. Some crops, notably fruit trees such as the apple, complete fruiting during the warm season and can withstand below-zero temperatures in the winter. Others have a hardy underground structure in the form of roots, bulbs, tubers, or rhizomes which maintain life while the upper parts of the plant die down during the winter. Annuals simply complete their full life cycle during the growing season, their hardy seeds being the medium for renewed life. Winter freezing can hardly be prevented in climates where it is the normal course of events. Crops adapted to the seasonal cycle of temperature must be selected. Hothouses are economically feasible for only certain high-value crops such as vegetables or flowers.

FROST

The greatest agricultural risk in connection with low temperatures is the threat of unseasonable frosts. Two kinds of frosts may be distinguished: (1) advection, or air mass, frost, which results when the temperature at the surface in an air mass is below freezing; and (2) radiation frost, which occurs on clear nights with a temperature inversion and usually results in formation of ice crystals on cold objects. The former, sometimes called *black frosts,* are more properly termed freezes. While air mass freezes are common in winter in middle and high latitudes, they are an agricultural problem primarily in relation to specific crops which are limited in their winter-hardiness. Actual plant damage may be the result of alternate freezing and thawing, frost-heave in the soil, or desiccation. Winter wheat, for example, can withstand freezing temperatures but is *winter-killed* if the roots are disturbed too much by frost-heave. If unusually severe, an air mass freeze can be disastrous in winter, but it creates a special hazard when it occurs in early autumn before plants have made the necessary physiological adjustments, or in late spring when field crops are in the seedling stage and trees and shrubs are budding or blossoming. In the subtropics a severe freeze is regarded as unseasonable at any time of year. Millions of dollars in crop damage have been sustained in subtropical climates as a result of cold air masses. The citrus industry is particularly vulnerable. Because the temperature of air masses cannot be controlled on a large scale, not much can be done to forestall the general hazard attending air mass freezes. High-value crops may justify the use of covers or mulches which reduce the loss of soil heat. Expensive shrubbery and small trees are often wrapped to prevent freezing and other types of damage. In spring, hotbeds and coverings can be provided to protect seedlings against moderate freezes. For production of most field crops, the only satisfactory solution to the problem of freezing is to avoid it as far as possible by planting after the danger is past in spring, and by selecting varieties which will mature before the renewal of the hazard in autumn. Even then the farmer may be taking a calculated risk.

Damage due to radiation frost differs from freeze damage in degree and in its spotty occurrence. Plants which are killed by a general freeze may be only partially damaged by frost, although the economic effects can be just as great. An entire fruit or berry crop may be wiped out by a single hard frost although the plants themselves are not necessarily killed. The hazard is greatest during critical stages of growth; for warm-climate crops, this means the entire growing season. Germinating seeds are not often affected by surface frost, but young seedlings may be killed unless they are of frost-hardy varieties. Crops like potatoes, tomatoes, and melons are vulnerable right up to maturity. The flowering stage is a critical period for most crops of field and orchard. In the spring-wheat belts of North America and Siberia, late summer frosts are destructive even after the kernels of grain have begun to form. Frosty nights followed by warm, sunny days produce a *sun scald* on orchard

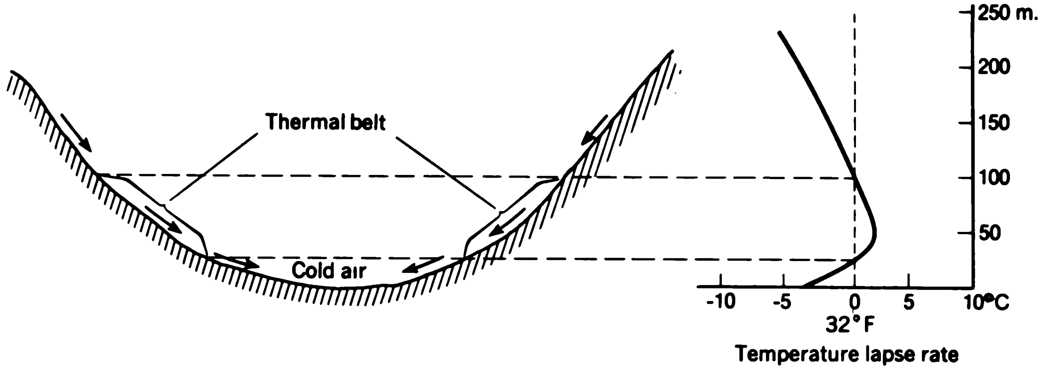


Figure 12.1
Thermal belt characteristic of some mid-latitude valleys.

fruits, considerably reducing their value. Tree trunks are sometimes affected by the alternate freezing and warm sunshine also; orchardists whitewash the trunks to decrease the absorption of heat during the day. In this way the range of temperature experienced by the bark is reduced.

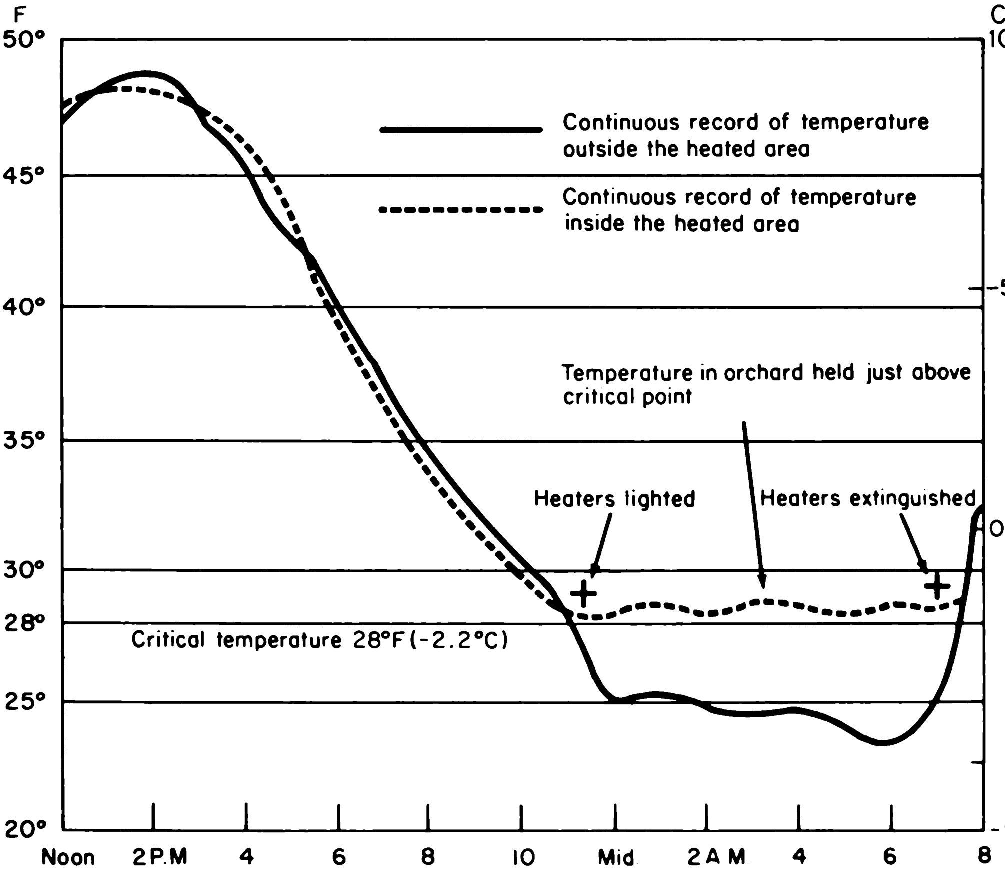
Prevention of crop damage due to radiation frosts is more feasible than thwarting a general freeze. Preventive measures are based upon a knowledge of the conditions for frost. These conditions are (1) a prevailing stable air mass with cool surface temperatures, (2) clear skies to allow loss of heat by radiation, (3) little or no surface wind to mix the cold air near the ground with the warmer air above, (4) a relatively high dew point temperature, and (5) topographic features which induce drainage of cold air into depressions. It is this latter factor which accounts for the spotty distribution of frost. Whether or not frost crystals will form on plant surfaces depends on the dew point of the air. If cooling takes place until the dew point is reached at a temperature above freezing, the formation of dew releases latent heat and retards further cooling. Below 0°C frost will form. This process also releases latent heat, which explains why light frosts of short duration may cause surprisingly little damage. Evaporation of water from soil or plants, especially after a previous rain, also serves to reduce the temperature at ground level, and it is frequently a contributory cause of frost.

As with freezes, a logical adjustment to frosts is to avoid them. Selection of frost-hardy plants is one way. Another is to grow delicate crops on slopes above the level to which below-freezing air is likely to extend or where air drainage or winds can be depended upon to stir the air throughout the night. North-south valleys are in the shade earlier in the evening than those which open to the west and inversions therefore develop sooner and have a longer period in which to concentrate freezing air in the valley bottom. In mid-latitudes mountain slopes frequently exhibit a *thermal belt* of maximum nighttime temperatures between fairly well-defined levels. (See Figure 12.1.) Below the belt, frosts associated with radiation and air drainage are a hazard; above the upper limit, low temperatures are related to the normal lapse rate and are negative factors in plant development by day as well as at night.

FROST PREVENTION

Direct frost prevention measures are largely aimed at breaking up the inversion which accompanies intense nighttime radiation. This is accomplished by stirring the air, heating it, providing a protective blanket of smoke, or by any combination of these. Obviously, the methods will work only in stable air. For many years smudge pots have been used to combat radiation inversions in citrus groves. Cheap oil or any fuel may be used, the objective being to create some heat at tree level and to form a dense pall of smoke which spreads out over the surrounding area and reduces the loss of heat by radiation in much the same way that a cloud cover does. In addition to raising slightly the temperature of the surface

Figure 12.2
Temperatures inside and outside an area using orchard heaters to prevent frost damage. (NOAA)



air, heat from the burners creates convection currents and brings about the desired mixing of air in the inversion layer. Of course, the soot settles on fruit trees and everything else in the vicinity, creating a problem in itself. By day a smoke cover retards warming at the surface. More efficient heaters which give off less smoke are now more common, and, although they do not provide the smoke blanket, the direct heating and resulting convective activity often successfully alleviate the frost danger. (See Figure 12.2.)

Another method of mixing stable surface air is by means of huge fans, usually operated by electric or gas-driven motors. Fans cannot prevent damage from an invading air mass with freezing temperatures, whereas heaters may at least reduce the effect of the freeze if it is not too severe. (See Figure 12.3.) Airplanes and helicopters have also been used to fan the air above crops, but this is an expensive procedure usually reserved for emergencies. Frost damage to wheat fields in northern states has been successfully prevented by the use of airplanes. The best-known frost prevention activities are those in the citrus areas of

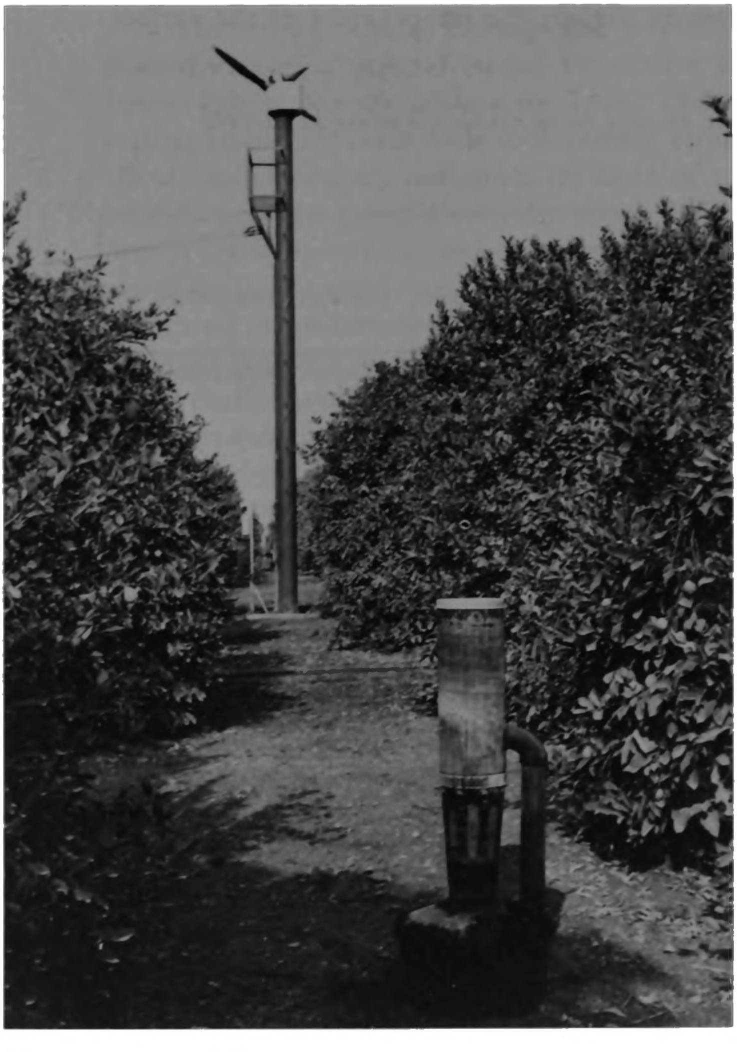


Figure 12.3
Fan and heater for frost prevention in a lemon grove. (Sunkist Photo)

California and Florida, where virtually every grove is equipped with thermometers and heaters or fans (or both). But these methods are widely employed in fruit orchards throughout the country, especially at the time of spring flowering. Where only a light frost is expected, water sprayed over fruit blossoms has proven effective in ameliorating the hazard. The water droplets tend to produce a protective cover of fog and maintain the temperature at or near the freezing point. In orchards and vineyards in Europe fine sprays have achieved protection against temperatures as low as −5°C. Experiments using artificial fogs composed of water droplets and emulsifying agents also show promise of providing a blanket to reduce heat loss. Orchards are a long-term investment, and expensive measures often prove economically sound. One heavy frost can mean the loss of a whole year's crop. Growers of winter or early vegetables can hardly afford to combat chance frosts by such methods, however. Even after a hard spring freeze, vegetables can usually be replanted; if it is too late for one variety, another can be substituted so that the financial loss is not complete. Tender plants are commonly covered with paper or straw for frost protection during the early stages of growth; this is an extension of the principle of cold-frame covers for seedlings. Some plants can be successfully covered with soil by plowing if there is advance warning, although the covering and uncovering processes cause a certain amount of damage. Flooding with water has also been employed as an emergency frost-prevention measure. Cranberry bogs are often saved from frost in this way. In certain circumstances, proper drainage of surface and soil waters can be equally important in frost prevention as a result of the reduction of evaporation. Cultivation practices, such as weeding and mulching, also can inhibit frost.

In order for frost-prevention methods to be effective, it is necessary to have advance warning of the danger. Forecasting of general freezing temperatures associated with a cold air mass is based upon analysis of air masses and predictions of their movements and modifications in local areas. While this is by no means a simple task, radiation frost forecasts are even more difficult because of the great local differences in the frost hazard. In the United States federal-state cooperation has made frost warnings especially valuable in the citrus-producing areas, where nightly reports are broadcast during the winter. Detailed forecasts of radiation frost are not feasible, but warnings of potential danger can be made. In a given locality the threat of damaging frost must be interpreted in relation to soil conditions and heat storage during the previous period, the stage of crop growth, topographic influences, the dew point, and wind, as well as the expected degree and duration of minimum temperatures below freezing. It is as important to know that frost will not occur as to know that it will, if the needless expense of preventive measures is to be avoided. Because of the spotty nature of radiation frost, every efficient farmer who faces a frost hazard has his property equipped with instruments to register local temperature, humidity, and wind so that he can determine the degree of danger for all parts of his orchard or cropland. He is alerted by frost warnings for his vicinity, but the final decision to light heaters or to begin other procedures rests on purely local conditions.

THE FROST-FREE SEASON

Climatic records usually derive the frost-free season from the number of days during which the temperature is continuously above 0°C, but for agricultural purposes the period between the last killing frost of spring and the first killing frost of autumn is a more useful index. The effects of freezing temperatures on the principal crops of a locality are the criteria for determining a killing frost, although crops vary in their frost-hardiness.

The dependability of a frost-free season is obviously important in agricultural planning. It is the occasional unexpected frost, especially in late spring or early autumn, that usually causes the greatest damage. Thus, the frost-free season that can be expected in a certain percentage of years offers a basis for choosing crops and calculating risk. For many crops it is better to operate well within the mean frost-free period to avoid risks that attend departures from the normal.

In general, the frost-free season decreases with an increase in latitude. Large areas within the tropics experience no frost except at high altitudes. The other extreme is at the poles, where there is no frost-free season; in the mid-latitudes numerous factors combine to produce wide variations both regionally and from year to year. For example, along windward coasts in the westerlies the frost-free period is longer than in continental interiors or on east coasts in the same latitude. In mountainous regions the seasonal regime of frost is highly complex.

The frost-free period is often equated loosely with the growing season, but if the latter is to have any practical meaning it must incorporate all of the climatic variables that affect crop growth and productivity. A previous section has reviewed temperature effects. In certain climates, especially in the tropics, the length of a growing season may be determined more by the availability of water to meet the needs of potential evapotranspiration than by temperature. The growing season of the dry summer subtropics clearly differs in character from that of the humid subtropics. At high latitudes the length of the daylight period and in-

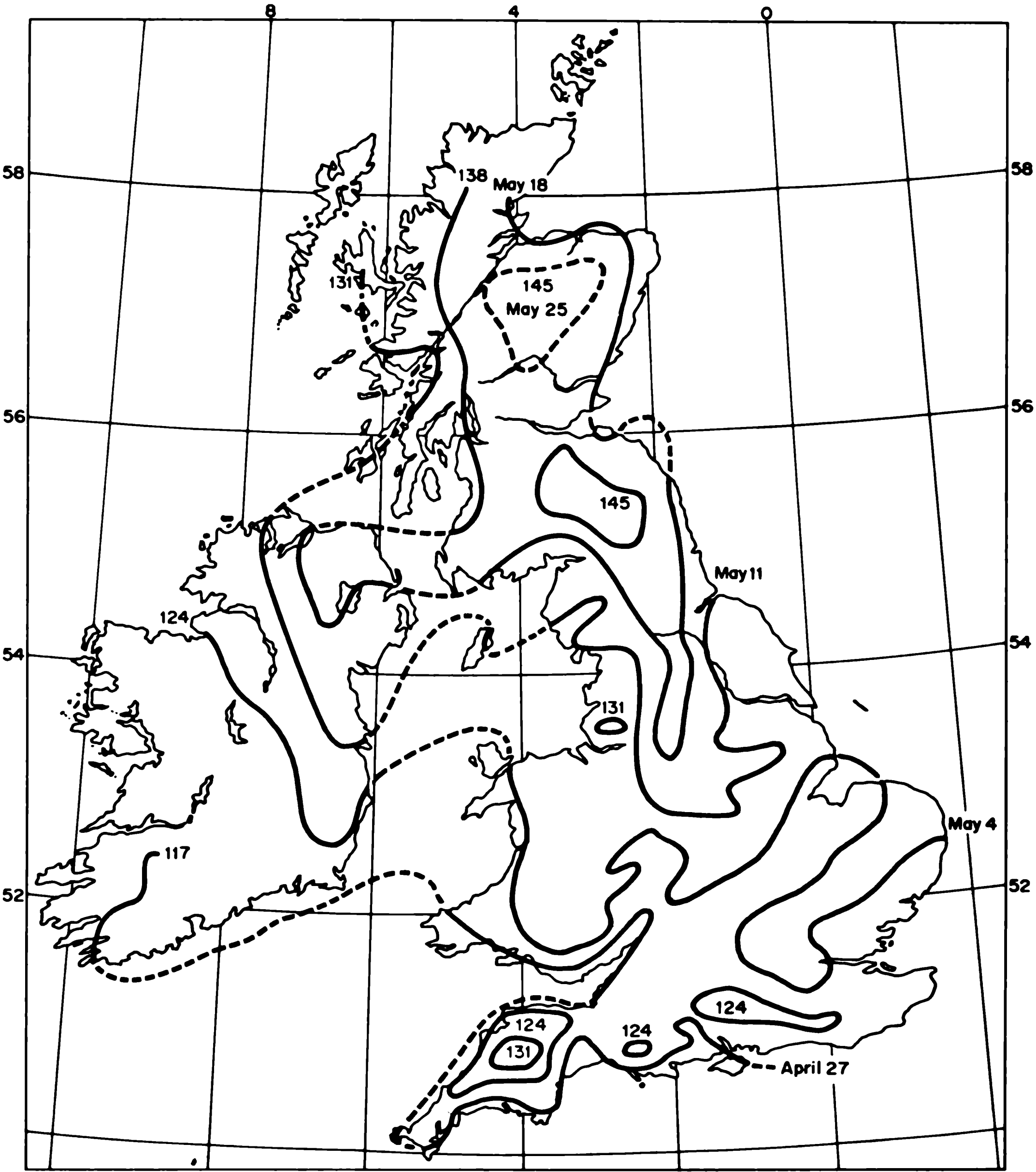


Figure 12.4
Isophenes of mean flowering dates, British Isles. Based on 12 selected plants for the period 1891–1925. (After Bush)

a function of climate, heat and light being the most important factors. A close relationship exists, therefore, between plant phenology and both latitude and altitude. It is well recognized that "spring moves northward" in the Northern Hemisphere and that "autumn moves southward." The frost-free season is progressively shorter with increasing latitude. In far northern latitudes the longer days in summer compensate for cooler temperatures to some extent so that certain crops can mature in approximately the same length of time as in mid-latitudes. Figure 12.5

tensity of summer insolation compensate in large measure for the short frost-free season. The concept of the growing season thus requires careful definition for each crop in terms of its total climatic needs, including freedom from hail, violent winds, or other destructive phenomena.

PHENOLOGY

Although temperature extremes that may damage crops are obvious climatic phenomena which affect agriculture, the rate at which plants develop in their various stages of growth is of equal importance in determining climatic limits of economic crop production. *Phenology* is the science which relates climate to periodic events in plant and animal life. Phenological data for crops include such facts as dates of planting, germination, and emergence of seeds; dates of budding, flowering, and ripening; and dates of harvest. These depend on climatic conditions preceding each event as well as on climatic factors at the time of the event. Their specific relationship to the climatic elements is not fully understood, but they nevertheless represent observed facts of plant growth that can be put to practical use. Generations of farmers have kept phenological diaries listing the dates of observed development of natural vegetation and crops as well as facts concerning the periodic reactions of birds and animals to climate and the seasons. Such information, recorded or not, has been the chief basis for denoting "signs of spring," "signs of autumn," and so on. When correlated with climatic observations, it aids the study of agroclimatic relationships, often revealing the influences of latitude, altitude, prevailing winds, or other factors. Maps organize phenological data by using *isophenes* to connect places at which a phenological event took place on the same date. (See Figure 12.4.) For example, the winter wheat harvest begins in Texas in June and gradually proceeds northward to Nebraska by early July. The difference in dates of significant phases of plant development over a given distance is the *phenological gradient*. Phenological gradients of microclimatic proportions are common perpendicular to garden walls, fence rows, or hedges. On the sunny side, plants are likely to develop faster close to a wall that reflects heat. If shaded by the wall, however, the same variety may mature later. Similarly, differences occur on opposite exposures of hills or mountains, although complications are introduced if the higher temperatures of sunny slopes are accompanied by a moisture deficiency.

The rate of development of a plant variety is the resultant of all the environmental factors: climatic, physiographic, edaphic, and biotic. For a particular field under standard cultivation practices, it is primarily shows graphically the phenology of the *Marquis* variety of wheat in several North American locations. The total time required from seeding to ripening of the wheat was 87 days at Fairbanks, Alaska (lat. 64° N), and 90 days at Lincoln, Nebraska (lat. 41° N). Obviously factors other than monthly temperature averages entered into the development of the wheat plants. Fairbanks has a mean July temperature of 15°C, whereas the comparable value for Lincoln is 26°C. The longer periods of daylight aided the maturity of the crop at the northerly latitudes and offset the lower temperatures. The importance of light in crop development also has been demonstrated in laboratory experiments.

At a given location the period between planting and harvesting is not a specific number of calendar days but rather a summation of energy units, which may be represented as *degree-days*. The duration of a certain temperature is quite as important as temperature averages. A degree-day for a given crop is defined as a day on which the mean daily temperature is one degree above the *zero temperature* (that is, the minimum temperature for growth) of the plant. Some representative zero temperatures are:

Spring wheat	0–4°C (depending on variety)
Oats	6°C
Field corn	12–14°C
Sweet corn	10°C
Potatoes	7°C
Peas	4°C
Cotton	17–18°C

Although it has proven to be a useful guide in agriculture, the degree-day concept has shortcomings. It does not differentiate between such combinations as warm spring—cool summer and cold spring—hot summer; it makes no allowance for unfavorably high temperatures nor for diurnal temperature ranges; and it neglects specific temperature values or trends that may be essential at various stages of crop development. During germination, for example, soil temperature at the seed depth is more significant than air temperature. Energy units expressed in terms of potential evapotranspiration or potential photosynthesis offer possible improvements in application of the fundamental concept to plants after they have emerged.

The time required to achieve maturity is also a function of the length of day, or *photoperiod*. In general a crop planted early in the spring requires more calendar days to mature than the same crop planted later. In mid-latitudes the days are shorter in spring and temperatures are lower. For efficiency in the use of labor and equipment at planting and harvesting times it is expedient to make successive plantings of a crop

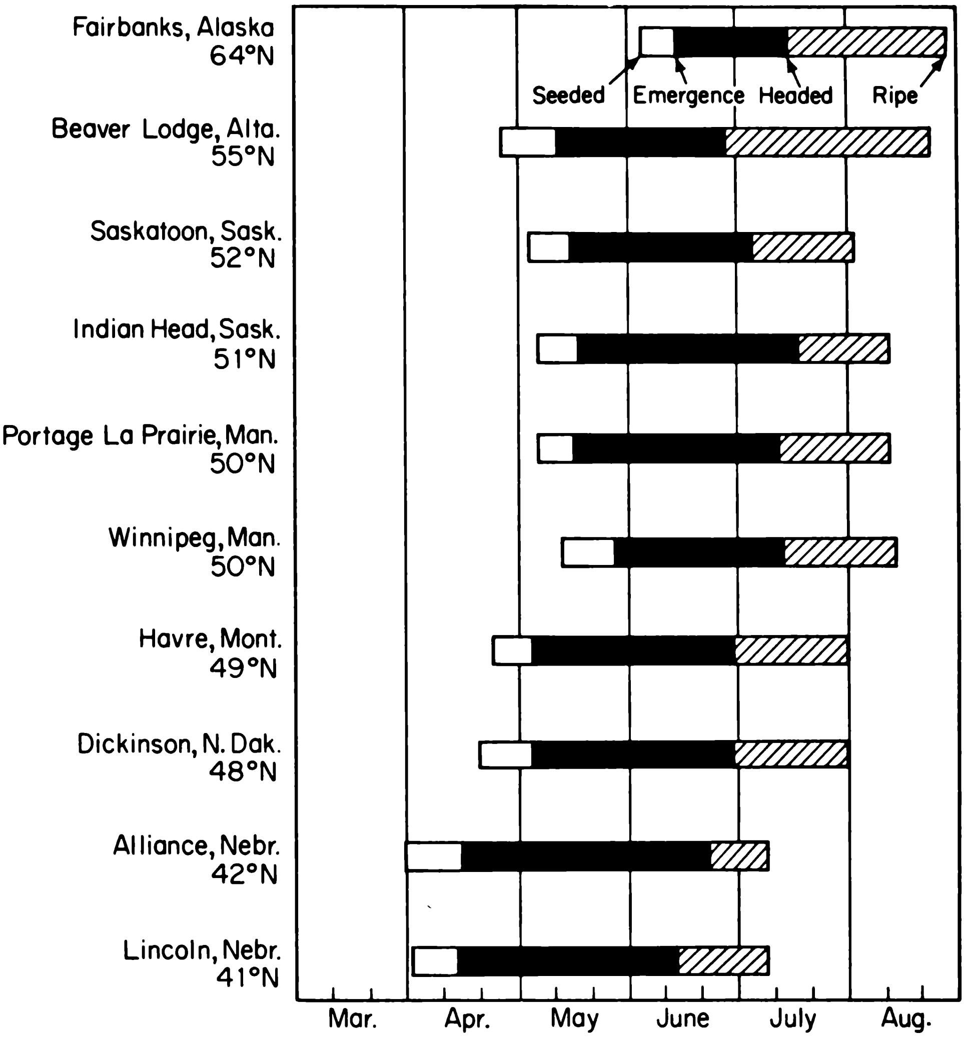


Figure 12.5
Chart of the phenology of Marquis *wheat at selected North American locations.* (From M. Y. Nuttonson, "Some Preliminary Observations of Phenological Data as a Tool in the Study of Photoperiodic and Thermal Requirements of Various Plant Material," in A. E. Murneek, R. O. Whyte, et al., *Vernalization and Photoperiodism.* Copyright 1948, The Ronald Press Company)

so that the ensuing harvest can be spread across several days or weeks. In this way all the facilities for harvesting, processing, and marketing can be kept to a minimum and yet can be used for a maximum time each season. This is especially significant for vegetable crops that must be harvested immediately at maturity in order to maintain quality. Let us consider green peas as an example. Table 12.1 gives the phenological data for Alaska peas planted on different dates at College Park, Maryland, in 1926. Note that the peas planted on March 29 took 74 days to mature to the harvesting stage, whereas those planted on May 8 required only 51 days. The average length of the photoperiod was 14 hours during the entire period for the first planting and 14.6 hours for the last, and average daily temperatures naturally increased as the season progressed.

TABLE 12.1
Phenology of Alaska Peas at College Park, Maryland, 1926*

Date planted	*Date emerged*	*Date first blossom*	*Date of harvest*	*Total days*
Mar. 29	Apr. 14	May 13	June 11	74
Apr. 3	Apr. 16	May 14	June 11	69
Apr. 9	Apr. 21	May 17	June 14	66
Apr. 16	Apr. 28	May 21	June 17	63
Apr. 24	May 4	May 27	June 22	61
Apr. 30	May 8	May 30	June 22	54
May 8	May 16	June 6	June 28	51

*Data selected from Victor R. Boswell, "Factors Influencing Yield and Quality of Peas," in *Biophysical and Biochemical Studies,* University of Maryland Agricultural Experiment Station, Bulletin No. 306 (College Park: March, 1929).

Careful studies of the phenology of crops over a number of years make it possible to establish planting schedules that are best adapted to local climate, availability of labor, and market demands. Sometimes it is also possible to avoid the periods of maximum hazard from insects, diseases, or seasonal weather phenomena. In irrigated areas phenological data can aid in determining the best time to plant in order to make the least demand on water supplies.

THE MOISTURE FACTOR

Within rather wide temperature limits moisture is more important than any other environmental factor in crop production. There are optimum moisture conditions for crop development just as there are optimum temperature conditions. Because crop plants obtain their water supplies primarily through their root system, maintenance of soil moisture is the most compelling problem in agriculture. Excessive amounts of water in the soil alter various chemical and biological processes, limiting the amount of oxygen and increasing the formation of compounds that are toxic to plant roots. The underlying cause of inadequate soil aeration may be poor vertical drainage as well as excessive rainfall. Therefore, conditions can be improved to some extent by drainage practices. On the other hand, a high rate of percolation of water through the soil tends to remove plant nutrients and inhibit normal plant growth. Cover crops and addition of humus to the soil help to alleviate this problem. Heavy rainfall may directly damage plants or interfere with flowering and pollination. By packing the surface soil, deluges make difficult the emergence of tender seedlings. Small grains are often beaten down, or *lodged,* by rain, and this makes harvest difficult and promotes spoilage and disease. The effect of rain on the harvesting and storage of grain and hay crops is a common problem. Special labor-consuming methods are employed to speed up drying and curing of hay to avoid rotting. Artificial drying of both forage and grain has been adopted as a method of combating damp conditions at harvest time. Wet weather during the later stages of maturity and harvest of cotton causes losses of both seed and fiber, often by providing favorable conditions for disease.

Snow and freezing rain are threats to wintering plants. The sheer weight of ice and snow may be sufficient to break limbs on trees and shrubs. A thick ice cover on the ground tends to produce suffocation of crop plants such as winter wheat. Hail is a special case of "excessive moisture" which causes direct damage to plants. Locally it may be a disaster, although over large areas it is a minor hazard compared to drought or the effects of low temperatures. The degree of damage depends on the stage of growth of the crop and upon the intensity of the hailstorm. Because hail is most common in the warm season, it frequently catches crops at a critical stage, pounding young plants into the ground, shredding leaves, or shattering flowers and seed heads. Figure 12.6 is a map of the average annual number of days on which hail occurs in the United States. The areas in which hail normally causes the most damage are in eastern Wyoming, eastern Colorado, western portions of Nebraska, Kansas, and Oklahoma, and in northern Texas. Many farmers in these and adjacent areas carry hail insurance on their crops. Premium rates are determined on the basis of weather records and hail damage records.

DROUGHT

Under natural conditions excessive moisture is far less an agricultural problem than is drought. *Drought* is the deficit that results when the

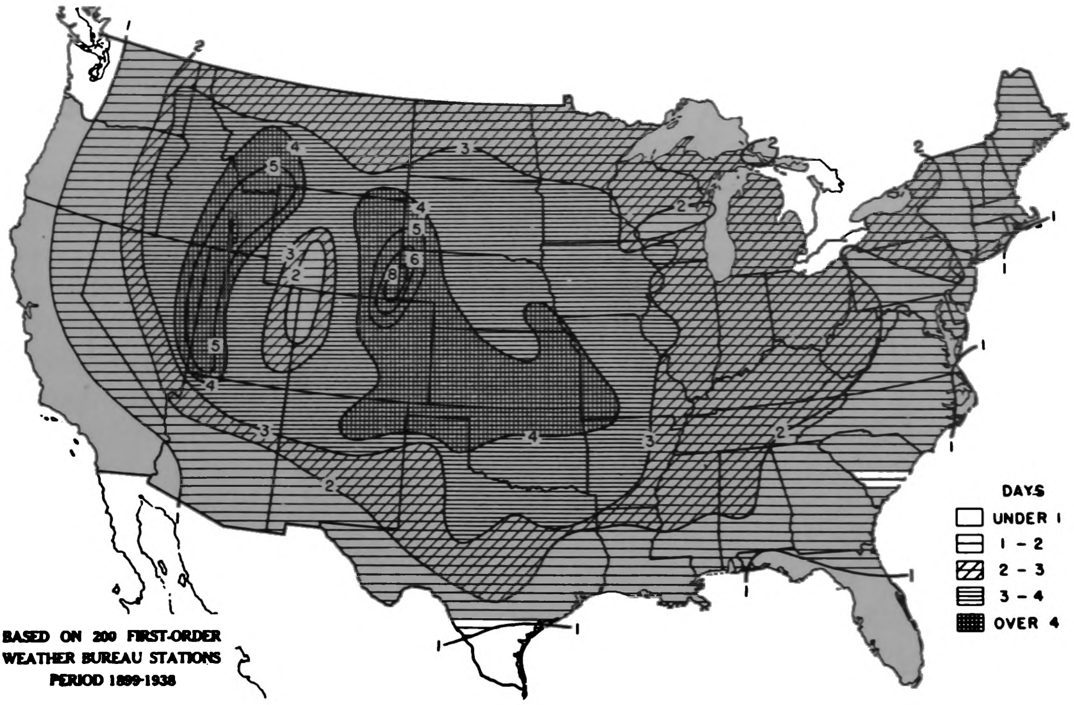


Figure 12.6
Average annual number of days with hail in the contiguous United States. (From *Climate and Man, 1941 Yearbook of Agriculture*)

amount of water available in the soil is insufficient to meet the demands of potential evapotranspiration. Three classes of drought may be differentiated: (1) permanent drought associated with arid climates; (2) seasonal drought, which occurs in climates with distinct annual periods of dry weather; and (3) drought due to precipitation variability. In every case, the underlying cause of droughts is insufficient rainfall, although any factor that increases water need tends to aid in causing drought. Low relative humidity, wind, and high temperatures are contributory factors because they lead to increased evapotranspiration. Soils which lose their moisture rapidly by evaporation or drainage also augment drought. Drought does not begin with the onset of a dry spell; it occurs when plants are inadequately supplied with moisture from the soil. Thus, crops growing on soils which have a high capacity for holding water are less susceptible to short periods of dry weather. Land-use practices which tend to increase runoff decrease vital soil moisture storage accordingly.

With these facts in mind, it is obvious that the incidence of drought cannot be determined alone from a map of average precipitation. In addition to the average amount, the seasonal distribution, dependability, intensity, and the form of precipitation must also be known. Furthermore, different crop plants have different moisture requirements. Ultimately, then, drought must be defined in terms of the *water need* of a particular

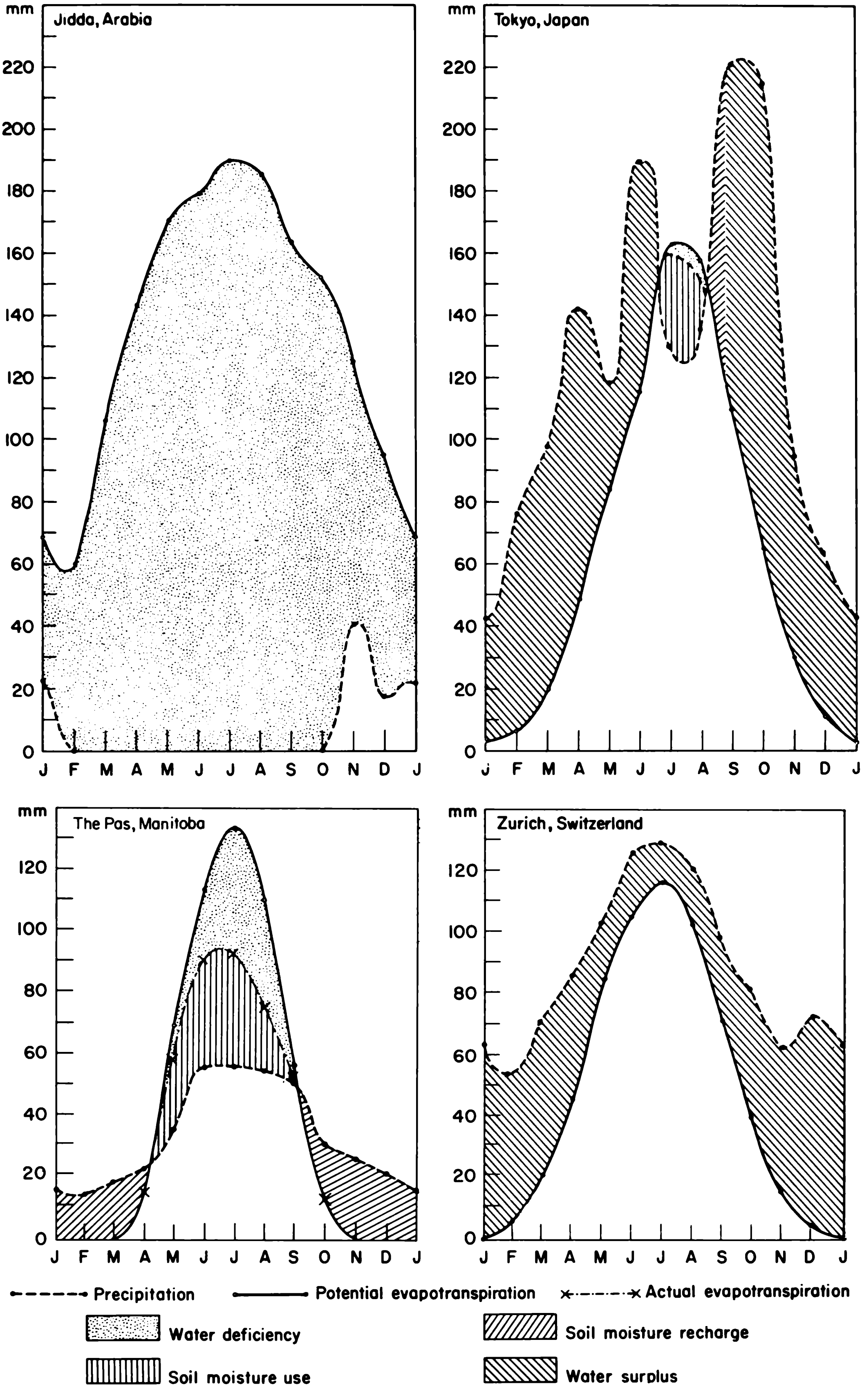


Figure 12.7
Water budget graphs for selected stations. (After method of Thornthwaite, based on data computed by C. W. Thornthwaite Associates, Laboratory of Climatology)

crop growing under a specific combination of environmental conditions. If the minimum water need is not met for these conditions the plants do not develop and mature properly.

Some climates have rainfall equal to or greater than the potential evapotranspiration in all months. The water-budget graph for Zurich, Switzerland, shown in Figure 12.7 (based on computed potential evapotranspiration) is representative of this. For most areas the monthly precipitation does not regularly exceed water need, however, and for part or all of the year there is a moisture deficiency. In the mid-latitudes the deficiency occurs in summer, when higher temperatures increase the rate of evapotranspiration. In the graph of Tokyo (Figure 12.7) note that during the first half of the year precipitation is in excess of water need. In the early summer, crops must depend on stored soil moisture and actually may suffer from a small moisture deficit. With the seasonal decline of temperature and the approach of the autumn rainfall maximum there is a recharge of soil moisture and a return to conditions of water surplus. In climates having distinct summer maximums of precipitation there may, nevertheless, be summer water deficits. Such is the case at The Pas, Manitoba (see Figure 12.7). Although precipitation in the summer months is several times that in winter, the long summer days and higher temperatures create water need in excess of the amount provided by rainfall and stored soil moisture. Note that the meager winter precipitation never succeeds in charging the soil to its full capacity. A degree of drought thus prevails throughout the growing season.

There is a broad relationship between permanent and seasonal types of drought and the world pattern of climates. Arid climates experience permanent drought, as the water-budget graph for Jidda, Arabia (Figure 12.7) indicates. Climates with well-defined seasonal minimums of rainfall are likely to have corresponding seasonal droughts, although their economic significance depends to some extent on whether they coincide with the growing season. In the dry summer subtropics the dry season, unfortunately, comes in the warm months. In the wet-and-dry tropics winter is the dry season, and summer cropping can be carried on if soil moisture is built up in time. A delay in the onset of the rainy season can be disastrous, however. Some of the worst famines of southern Asia have resulted from failure of the summer monsoon to develop as early as usual. Variability of precipitation creates the greatest drought hazards. This is especially true in the semiarid climates, where a slight departure from the average may be the critical factor in crop failure. Even in humid climates there may be true drought which reduces the yield of crops, even though it does not wipe out harvests entirely. There are degrees of drought and consequently of drought damage; the effects of moderate drought are not always exhibited in such obvious forms as withering plant leaves but may appear instead in lowered quality or yield.

The water-budget graphs referred to above represent average conditions over a number of years. They cannot show the haphazard occurrences of dry weather which plague the farmer in normally wet seasons or in humid climates. Nor do they represent the year-to-year fluctuations in precipitation and therefore in the drought hazard. These are, nonetheless, significant to agriculture. Several years of abnormally dry climate are usually accompanied by higher-than-average temperatures which accentuate the drought conditions. Prolonged drought conditions serve to alter the pattern of agricultural land use on a major scale. In the past they have led to migrations from affected areas, for example, the exodus from the Great Plains in the 1930s.

COMBATING DROUGHT—IRRIGATION

As we have seen, drought is a condition where water need is in excess of available moisture. Prevention of drought damage to growing crops, then, is a matter of either (1) decreasing the water need of crops or (2) increasing the water supply, or possibly a combination of the two. Planting of crops that have low water demands helps reduce the water need. Cultivation practices which improve the soil structure and inhibit runoff are effective drought-prevention measures, although they have limitations. Weed control is especially important if the available water is to be used most effectively for crops, for weeds accelerate water loss by transpiration at the expense of soil moisture.

In subhumid and semiarid climates *dry farming* methods depend on the conservation and use of two—or sometimes three—years' rainfall for one year's crop. During the period when a field lies fallow it is cultivated to kill weeds and to create a soil structure that will retain as much moisture as possible. Thus, soil moisture storage accumulated during the fallow period supplements the meager rainfall in the crop season. In the case of seasonal droughts the planting schedule can sometimes be adjusted to permit maturity and harvest before the effects of the dry season become too great. This is possible only if the temperatures are high enough in the wetter season.

Wherever the water need of crops or grassland cannot be reduced to conform to the moisture supply, the only alternatives are to abandon agriculture or to provide water artificially. At best, the artificial stimulation of rainfall has rather narrow limitations for supplementing natural precipitation. Whatever else may be claimed for it, the fact is that it has

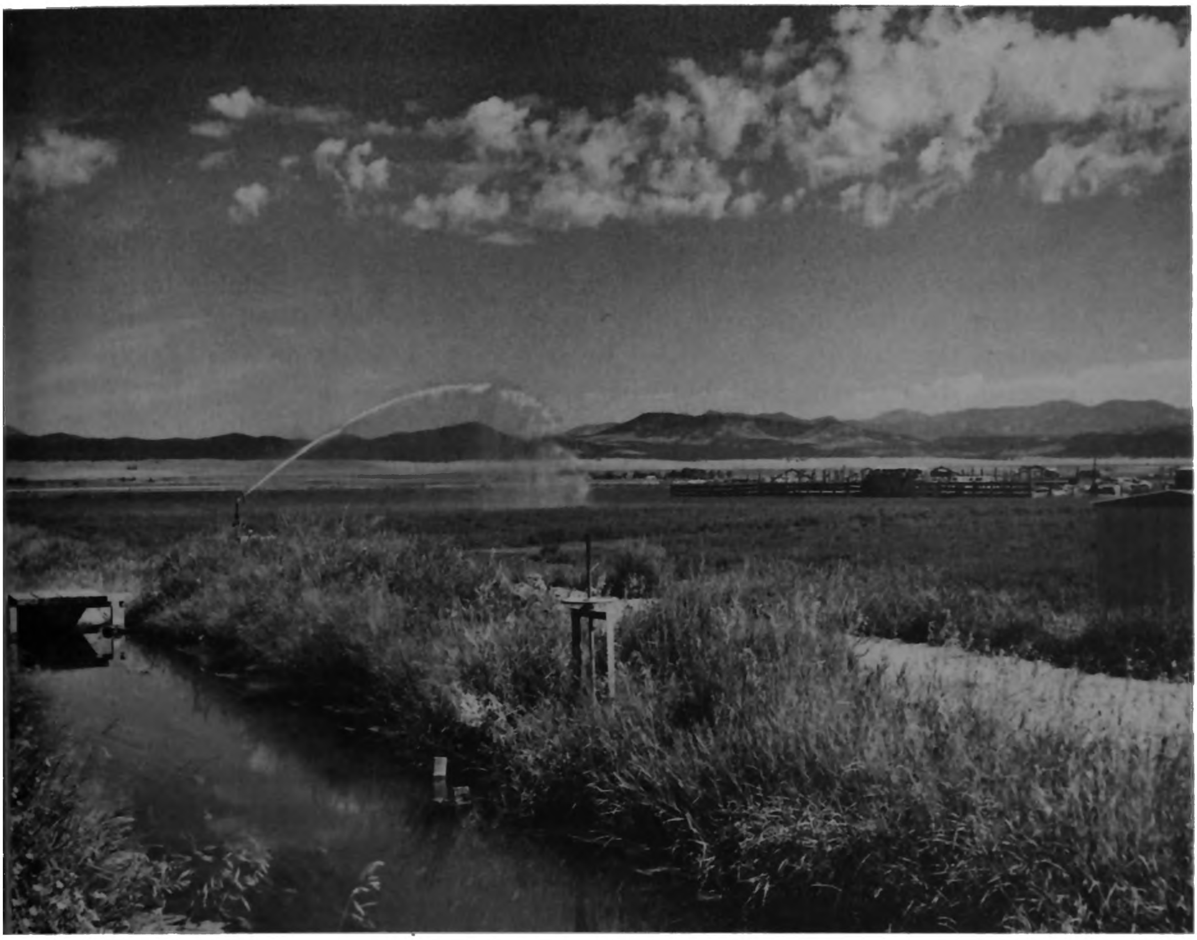


Figure 12.8
Sprinkler irrigation of alfalfa in western Montana. Evaporation claims part of the water before it can be effective in plant growth. The fractocumulus clouds are unlikely sources of precipitation. (Bureau of Reclamation, U.S. Department of the Interior)

not had any importance in combating drought in the past. On the other hand, irrigation is a widespread method for providing all or a part of the water need of crops. (See Figure 12.8.) In arid regions, or where cropping must be confined to a warm, dry season, agriculture is possible only with irrigation. In semiarid and subhumid climates, irrigation makes possible larger yields and a greater variety of crops. It also lengthens the period during which land can be used productively. In humid regions its main value is supplementary in times of drought. The chief limitations on irrigation are the availability of water from surface or ground water sources and the cost of getting it to the fields.

Within these limits, irrigation has the advantage that it can be regulated as an element in the water budget to meet the variable demands of different crops, different seasons, or chance droughts. Ranging between the obvious necessity for irrigation in deserts and the almost total absence of need for it in constantly wet climates are the variable circumstances that determine the amount of irrigation needed and when it should be applied. The total amount of water required for crop production is equal to what is used in transpiration and evaporation, plus what is lost by percolation below the root zone and unavoidable forms of waste incurred in the irrigation process or in rainfall runoff. The proportion that has to be provided by irrigation depends on rainfall. In general, irrigation is desirable whenever the soil moisture storage in the root zone drops to about 40 percent of capacity. In any case, to be effective, water must be applied before conditions reach the wilting point.

The principal advantage in using either soil moisture measurements or evapotranspiration as an index of irrigation needs lies in the resulting efficiency of water use. It is wasteful to irrigate before water is needed or to apply too much water. The absence of rainfall for several days may or may not be an indication of the need for irrigation. Temperature is likely to be a more important factor, since it directly affects evapotranspiration.

In summary, irrigation may be feasible in any climate where water need exceeds precipitation for a long enough period to reduce crop yields. Irrigation is man's best answer to drought. Where irrigation water is made available, temperature becomes the dominant climatic factor controlling crop distribution and yields.

CROPS AND WIND

Wind has its most important effects on crop production indirectly through the transport of moisture and heat in the air. Movement of air increases evapotranspiration, but the effect decreases with increasing wind speed and varies among plant species. Moderate turbulence promotes the consumption of carbon dioxide by photosynthesis. Wind may speed the chilling of plants or, on occasion, prevent frost by disrupting a temperature inversion. The action of the wind as an agent in the dispersal of pollen and seeds is natural and necessary for native vegetation and may be helpful for certain crops, but it is detrimental when weed seeds are spread or when unwanted cross-fertilization of plants occurs. Continuously strong winds interfere with the pollination activities of insects. Direct mechanical effects are the breaking of plant structures, lodging of hay and cereals, or shattering of seed heads. Fruit and nut crops may be stripped from the trees in high winds. The unfavorable consequences of depletion of soil by any form of erosion should be obvious. Low plants are sometimes completely covered by windblown sand or dust;

abrasion of plant stems and leaves by sand particles is often associated with wind erosion. Along the shores of salt lakes and oceans, salts transported inland by the wind affect both plants and the soil.

Practices to avert the effects of wind on evapotranspiration are essentially those, including irrigation, employed to combat drought. Mechanical damage due to wind can be lessened somewhat by making use of natural or artificial shelter. Protected valleys and lee slopes are suitable for some types of crops which are easily damaged by wind. Windbreaks composed of trees, shrubs, hedges, or fences are widely used to protect both crops and animals from the wind. Some plants require only temporary protection which can be provided by screens or individual windbreaks. An example of the latter is the use of boards or metal cylinders to shelter young tomato plants. The best permanent windbreaks are rows of trees planted perpendicular to the prevailing winds. (See Figures 12.9 and 12.10.) Their moderating effect is felt for a distance equal to several times their height. Where winter snow is common, they have the added advantage of inducing snowdrifts over the adjacent fields. As a widespread measure to combat wind, they have some disadvantages, however. They reduce the area of cultivated land, compete for soil moisture, and may produce harmful shade. The first of these objections can be overcome if the species can be selected and managed so as to provide a compensation in the form of harvest from the windbreak. If too dense, windbreaks generate turbulence that may nullify their intended effect.

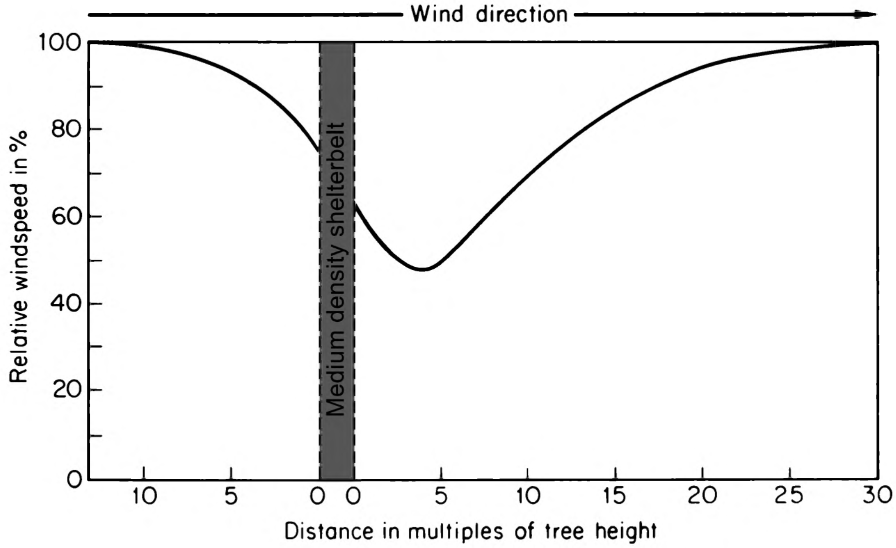


Figure 12.9
Relative wind speeds in the vicinity of a medium-density shelterbelt. (After Staple)

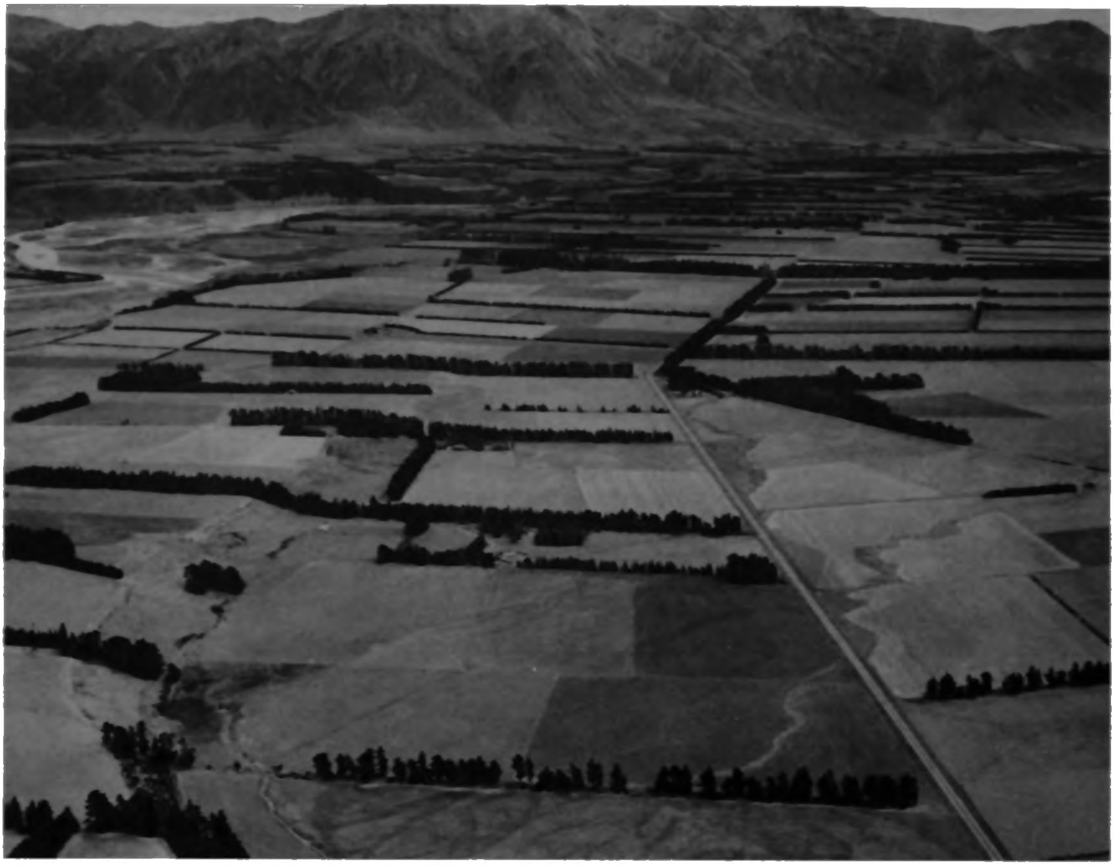


Figure 12.10
Shelterbelts on the Canterbury Plains, New Zealand. The rows of trees, mainly conifers, reduce the effects of a dry northwesterly foehn wind on soil, crops, animals, and farmsteads. Note the alignment perpendicular to the valley. (N.Z. Department of Scientific and Industrial Research)

CLIMATE AND FORESTRY

In modern forestry, timber is regarded as a crop and, as such, it is subject to the influences of weather and climate just as field crops are. The relationships between climate and forests are at least partially reciprocal. The distribution and rate of growth of forest species are influenced by climatic conditions, and forests, in turn, affect the climate on a local scale. The major effect of trees on climate is felt in the area they occupy and is roughly proportional to the density of the forest cover, but forests modify the climate of adjacent areas as well. Temperature averages within a forest are slightly lower than in adjacent open areas and the ranges are not so large.

In the mid-latitude lowlands the mean annual temperature within a forest is about 0.6C° lower than on the outside; at an elevation of 1,000 m the difference is 1C°. The greatest average difference is in summer, when it may be as much as 2C°; in winter it is only about 0.06C°. On the hottest summer days a forest may reduce the surface air temperature by more than 2.5C°, whereas in extremely cold winter weather the difference is only about 1C°.

In low-latitude forests the influence of a tree canopy on air temperature is even greater. In a dense forest the upper canopy shades the ground and acts as the primary absorbing surface during the day, thus retarding the rise of soil temperature. At night the canopy radiates heat more rapidly than the ground, which is slower to cool. The same principle holds true for seasonal variations. In spring and summer, the soil of the forest floor is slow to heat but in autumn and winter it remains slightly warmer than soil outside the forest, though the difference is less. Depending upon the density of the cover, forests may intercept up to 90 percent of the sunlight which is incident at the treetop level. Anyone who has tried to take a photograph inside a forest can attest to this. The reduction in light is the primary cause of paucity of low plants under a dense forest canopy such as is found in the tropical rain forest.

Clear evidence that total precipitation over a forest differs appreciably from that over adjacent areas is lacking. Rain gauges placed in open spaces within the forest usually indicate greater rainfall than outside the forest, but this is probably due to their being protected from the wind so that they collect a more accurate sample. Tree crowns intercept a part of the precipitation which falls over a forest, and much of it is evaporated

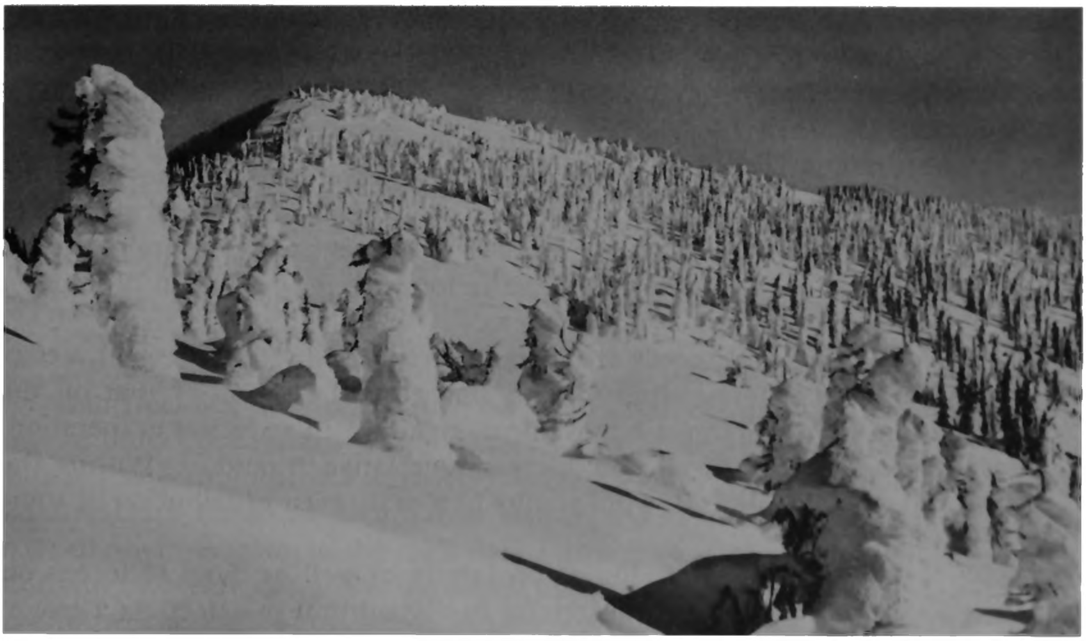


Figure 12.11
Intercepted snow in conifers on a mountain in northern Idaho. Sublimation and evaporation of melt water in the suspended snow reduce the runoff from the area. (U.S. Forest Service)

back into the air. The proportion thus withheld from the soil depends on the density of foliage and also on the duration and type of rainfall. A light rain of short duration may be almost entirely caught in the canopy of a dense forest, whereas in a driving rainstorm accompanied by wind most of the rain reaches the ground. Interception of advection fog may produce fog drip and enhance soil moisture. Snow collected in the crowns of trees may be shaken out by the win or it may melt and drip to the ground or flow down the trunks. On the other hand,a great deal of this potential soil moisture is lost by sublimation of the snow or by melting and evaporation (Figure 12.11). In general,coniferous forests intercept more precipitation than do hardwoods and deciduous species prevent comparatively little precipitation from reachin the ground in winter. By thinning or clear-cuttin portions of evergreen forest it is possible to reduce the interception of snow in treetops and admit more snow to the ground where it can augment runoff with smaller losses due to sublimation. Conversely, reforestation may actually result in decreased stream flow from a atershed. Increased transpiration as well as evaporation of water intercepted by the forest canopy reduce surpluses in the water budget Management of forested watersheds often entails decisions on the relative value of water and timber production.

Relative humidity is 3 to 10 percent higher within a forest on the average owing to the lower temperatures, lighter air moviment and transpiration from plants. Evaporation from the soil is considerably lessened as a result of the protective influence of the forest, and,if the ground is well covered with plant litter, it is reduced by one-half to two-thirds as compared to evaporation from soil in the open. During the growing season, trunks and branches have lower temperatures than the surronding air and they are further cooled by radiation at night with the result that dew or fog is formed. The movement of air from a forest to adjacent open country often carries fog with it and aids in preventing frost.

Surface wind speeds are markedly reduced by trees several meters inside a dense forest wind force is but a small fraction of that on the outside. By restricting air movement,trees aid in reducing evaporation, lowering temperatures, and increasing relative humidity. Within the forest, the reduction in wind results in a more even distribution of now cover.

The influences of climate on forests, as well as those of forests on climate, must be taken into account in silvicultural practices. As a group, forests species have higher moisture demands than do most other types of plants, and forests are therefore associated with humid climates. On a smaller scale, the variations in climate associated with local differences in slope, exposure, or altitude influence the distribution and rate of growth

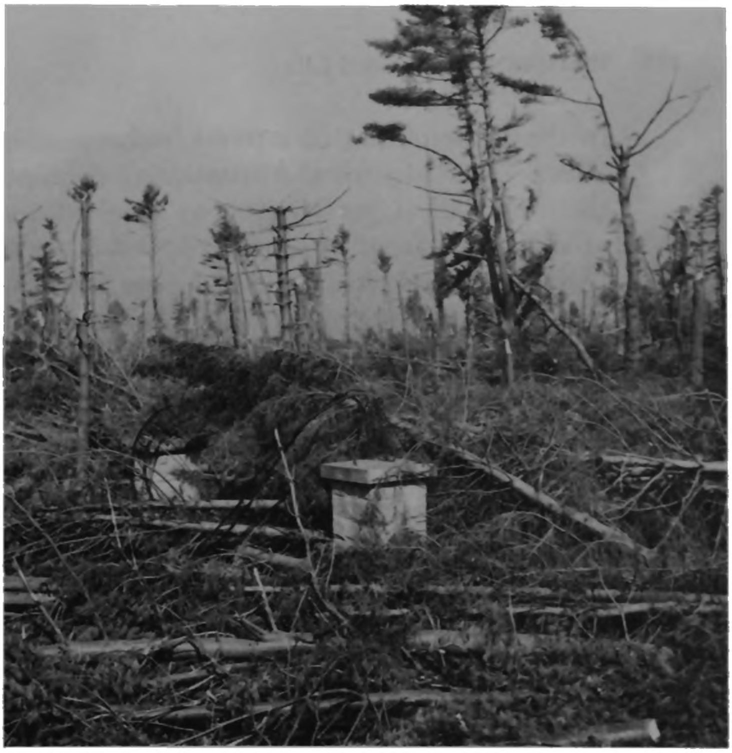


Figure 12.12
Hurricane damage in a New England forest.
(U.S. Forest Service)

of trees. (See Figure 10.7.) Individual tree species, like field crops, have optimum climatic requirements. Favorable conditions must be met by both the general atmospheric climate and the climate within the forest itself. In a climax forest dominated by one or more species, other trees equally adapted to the regional climate may not be able to gain a foothold because of shading or other environmental effects created by the existing forest. When burned or cut over, the climax forest species may be unable to regenerate without a favorable environment provided by intervening plant associations. Some species require protection from direct sunlight and the accompanying high soil temperatures when in the seedling stage but later become aggressive enough to crowd out trees that afforded the protection.

In the practice of scientific forestry it is possible to control the forest microclimate by logging methods, for example, thinning and burning, in order to maintain a desired forest association, whether it be the climax formation or not. Primarily, this entails control of the amount of light reaching the forest floor. Dense stands tend to develop tall, straight trees of small diameter as the trees fight upward for sunlight. Subsequent thinning, whether natural or artificial, allows more light and space for growth as the trees approach maturity. Selective logging of large trees often "releases" smaller trees of the same age so that they begin to grow much more rapidly. Too-heavy cutting of certain species may expose the trees to wind damage, however. Windfall is especially common along the margins of clear-cut areas in dense coniferous forests. (See Figure 12.12.)

In the perpetuation of a forest, reforestation, or the planting of a new forest, knowledge of the climatic environment which is best suited to growth at all stages is basic to most efficient management for timber production, watershed protection, and other forest values.

Since trees have a much longer life span than field crops, they are influenced much more by climatic fluctuation through the years. During a series of dry years their growth is limited; in wet periods they grow more rapidly. Extreme temperature fluctuations also influence growth; high average temperatures ordinarily accompany dry years and cool periods go with wetter climate. Increasing evidence of the fluctuations in climate during the life history of trees is being gained by inspection of their annual growth rings. This subject will be discussed in greater detail in Chapter 16 as it pertains to climatic changes and cycles.

FOREST-FIRE WEATHER

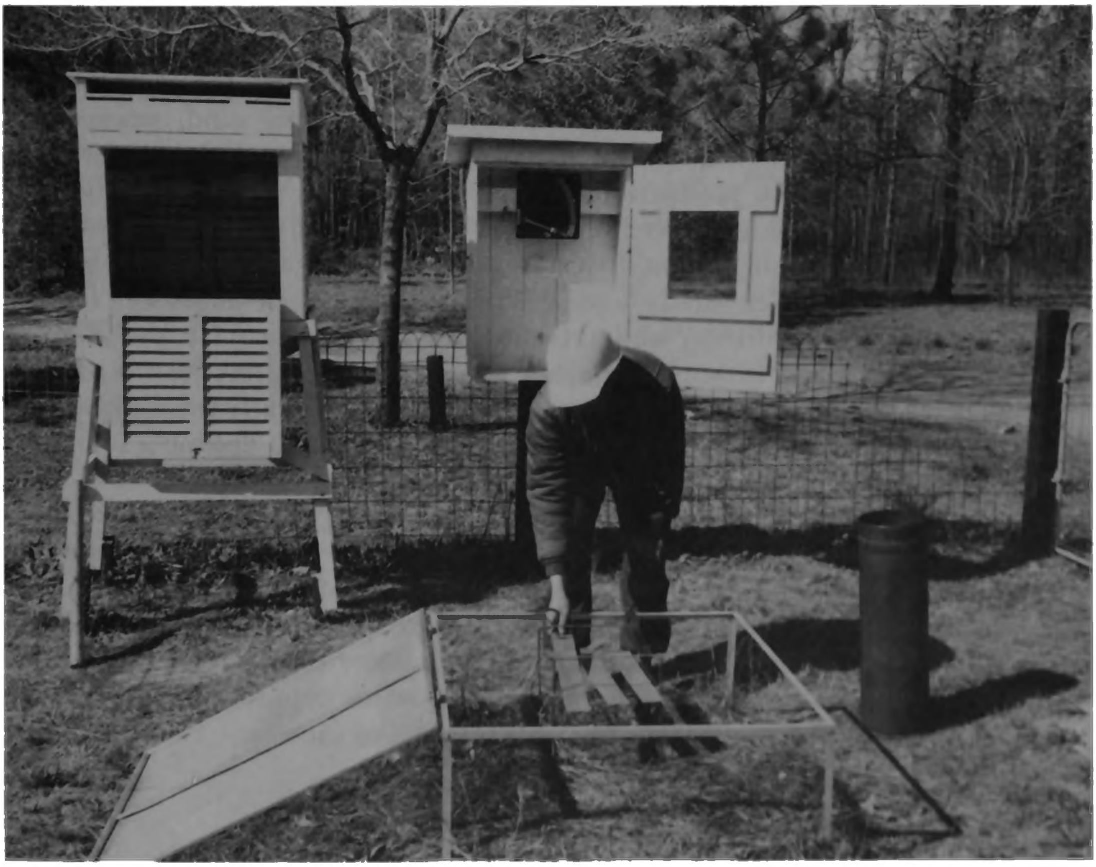
One of the greatest enemies of forests is fire, and forest-fire prevention and control constitute one of the most expensive, albeit necessary, phases of forestry management. Innumerable fires are directly caused by lightning. It has been estimated that lightning sets 5,000 to 10,000 fires annually in the forests of the western United States alone. Fortunately many of these are kept from spreading by rain. Others are fanned into disasters by high winds. Prevention of lightning-caused fires could conceivably begin with control of lightning. To whatever extent cloud-seeding techniques can moderate thunderstorms, this is a possibility. Forecasts of possible lightning storms and a knowledge of forest conditions resulting from previous weather can be valuable in determining the best techniques for arresting a fire. Because lightning strikes are more common on elevations, fires thus started are often much more difficult to reach than man-caused fires and their control calls for special methods and equipment.

The potential occurrence of forest fires and their rate of spread once started are both direct functions of weather conditions. The previous weather is quite as important as that currently prevailing, since it influences the degree of flammability of forest materials. Factors considered in determining the fire danger include relative humidity, temperature, wind speed and direction, precipitation, and condition of the vegetation and litter. Extreme fire danger results from long periods of hot, dry weather, especially when associated with stable air masses. Wind aids in drying out the forest and considerably increases the rate of spread of a fire. During the fire season, observations of the weather elements are taken at numerous stations operated by various public and private agencies. These stations employ standard instruments and usually have in addition a device for determining the moisture content of forest litter. This *fuel moisture indicator* consists of wooden sticks that are mounted in the open where they are freely exposed to the weather (Figure 12.13). They are periodically weighed on a scale which is calibrated to give an index of fuel moisture content. Analysis of observational data in the light of general forest conditions is combined with the regional weather forecast in preparing fire-weather forecasts and warnings. These reports are the basis for restrictions on logging and travel in forests and for the readying of fire-fighting crews. Mobile weather stations also give local information at the scene of a fire. Only with complete knowledge of weather conditions can forest fires be prevented or controlled effectively.

CLIMATIC FACTORS IN ANIMAL HUSBANDRY

Since domestic animals are highly dependent on the availability of feed, the climatic factors which influence the growth of pastures or feed crops

Figure 12.13
Fuel moisture indicators. The wooden indicators (foreground) are weighed periodically in the shelter to determine an index of forest fire danger. (U.S. Forest Service)



exert an indirect influence on livestock. The suitability of a particular breed of animal to a climate depends on the quality and quantity of feed which is available naturally or which can be grown in that climate quite as much as on the physiological adjustment of the breed to the climate. It is often feasible to transport feedstuffs into areas where grain or forage crops are inadequate, but this is subject to economic considerations. Generally it is satisfactory only where the demand for an animal product is very great or where there is a temporary shortage of feed because of climatic or other factors. Fluctuations in the output of animal products frequently result from variations in the feed supply rather than from the direct effects of climatic elements on the animals. Thus, there may be a decrease in milk production during a drought period because the amount and nutritive quality of pastures are impaired. Similarly, meat animals may not fatten in the normal length of time if unfavorable climate interferes with their feed supply. In general, animals are adapted to a wide range of climatic conditions provided that their feed and water requirements are met. There are, however, direct climatic effects on the normal body functions of animals. All breeds have optimum climatic requirements for the maximum of growth and development. When removed to a much different climate they do not ordinarily die as plants often do, but they may fall below minimum economic levels of production.

The climatic elements which affect livestock indirectly through the feed supply are those which influence plant growth or the spread of insects and diseases. Those which have direct effects are temperature, precipitation, relative humidity, atmospheric pressure, wind, storms, and light. Of these, temperature is the most important. High temperatures generally reduce production (see Figure 12.14). It has been found that dairy cows produce less milk under high temperatures, the optimum temperature being about 10°C. Hens produce larger eggs in winter and at high latitudes. Many animal breeds reduce their ingestion of feed under hot conditions and therefore do not produce fat and flesh so rapidly. Under extreme heat, they experience discomfort just as humans do. The reproductive capacities of animals are generally reduced by high temperatures. Mid-latitude breeds show a marked decline in fertility when relocated in tropical climates. The effect of extreme cold is to lower production also, for too much body energy is required to combat the cold. Where seasonal contrasts are great, most domestic animals adjust to winter by undergoing certain physiological changes, notably the growth of thicker coats of hair. Long exposure to cold, especially if it is accompanied by wind, may cause frostbite or death. Another problem associated with cold weather is the water supply, for animals will not satisfy their water requirements from ice or snow.

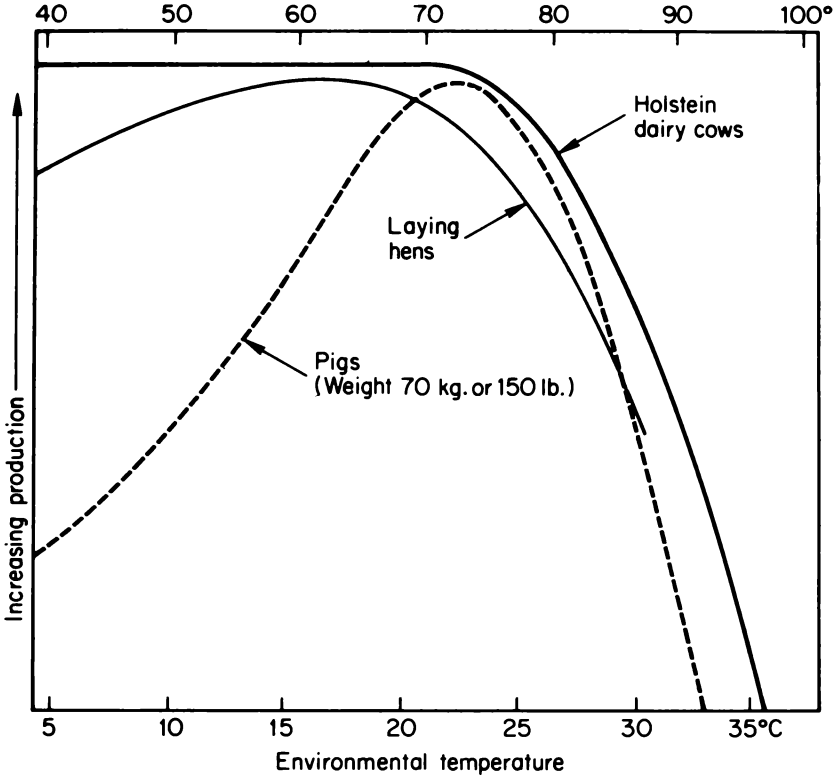


Figure 12.14
Production trends of three farm animals under varying environmental temperatures. (After Bond and Kelly)

Sunlight and the duration of daylight influence animals in several ways. Breeds of cattle and pigs with light-colored skins are occasionally sunburned under intense sunshine. The daily feeding period of most classes of livestock is determined in part by the length of day. Cattle normally graze more in shaded locations than in the sun and rest during the midday periods of highest temperature and strongest sunlight. Fertility is also affected by the duration of daylight, especially in poultry. Commercial poultry farmers commonly extend the day by means of artificial lighting in order to increase egg production.

The primary influence of precipitation on livestock is through its effect on feed. If it is associated with cold weather, it tends to accentuate the detrimental effects of the low temperatures. Freezing rain can cause much discomfort to say the least. Heavy snow makes moving and grazing difficult or impossible for range animals. Crusted snow or ice may cut their feet and a heavy cover of snow over small animals can cause suffocation. In blinding snowstorms, cattle and sheep often pile up against barriers, where they are trampled or die of suffocation, or they may stampede over precipices. Ranchers who use mountain pastures in summer are well advised to move their livestock to lower elevations before the time of possible autumn snowstorms. For the same reason, too-early grazing of mountain pastures in spring must be avoided.

High relative humidity, whether associated with abundant rainfall or not, influences respiration and perspiration in animals. Very dry air may cause discomfort, but it is ordinarily of less importance than high temperature in a drought provided that feed and water are available. Low relative humidity, wind, and high temperatures all increase the water requirements of animals. Even the camel can go but a few days without water under the extreme heat and dryness of the tropical deserts in summer. Wind may have good or bad effects, depending on wind speed and the accompanying temperatures. Moderate breezes ameliorate discomfort and ill effects caused by high temperatures. On the other hand, high winds increase drying, may fill the air with dust and sand, or intensify the impact of precipitation. In extreme cases, the force of the wind may be sufficient to blow down the shelters or buildings intended for protection. Moving cold air produces chilling much more rapidly than still cold air of the same temperature.

Changes in atmospheric pressure appear to have some effect on animals. The relatively large pressure differences accompanying changes in altitude are especially significant. For example, the fur-bearing chinchilla had to be "acclimated" to higher pressures in successive stages in order to transfer breeding stock successfully from the high altitudes in the South American Andes. Probably the lesser pressure changes which occur with cyclonic storms are felt to some degree by animals. Decreasing pressure may be one of the stimulants to intensified feeding prior to the arrival of storm centers or fronts.

The climatic elements are combined in storms and they do not act separately on animals. Wind, temperature, and precipitation are the three most important elements. In thunderstorms, lightning may be added to the hazards.

Many of the practices in animal husbandry designed to lessen the negative effects of climate depend on the use of natural or artificial shelters. Specific problems such as the control of temperature can be solved to some extent by heating and air conditioning. Experiments with dairy cattle in the tropics have shown that milk production can be increased by cooling stables with an air-conditioning system. Heating is widely employed in farrowing pens for sows and in brooder houses for poultry. Well-built barns and sheds help to conserve animal body heat in cold climates and in winter; in many instances this is sufficient. Overheating of buildings generally has a detrimental effect. Heating of dairy barns in winter results in reduced milk production. It is usually impractical to provide buildings for large numbers of range livestock. They are best driven to sheltered locations where feed and water are available. Windbreaks of trees and shrubs afford shelter to animals kept in the open.

Some animals gain relief from high temperatures if their coats are clipped. The outstanding example is sheep, which are normally shorn in spring. Cows and horses also benefit from clipping.

SELECTION AND BREEDING FOR CLIMATIC ADAPTATION

Where a suitable environment cannot be provided economically for a specific breed, the alternative is to select breeds or to breed new types that are adapted to prevailing conditions. Numerous examples might be cited of the adaptation of animal breeds to a particular set of climatic conditions. The Merino sheep is well suited to the semiarid climates of mid-latitudes. The English mutton breeds, on the other hand, do best in cool, humid climates. Northwest European dairy breeds such as the Holstein and Guernsey produce best in the mid-latitudes. The Jersey is considerably better adapted to warm climates than other common dairy breeds. As an extreme example, one can point to the reindeer, which has achieved a natural adaptation to the arctic environment. By selecting individuals that have, in some degree, the desired qualities of tolerance it is frequently possible to breed lines that are better adapted to a given range of climates.

Crossbreeding of animal types to combine the desirable qualities of the foundation stock has been successful among several classes of livestock. Since the heat tolerance of European cattle breeds is low, crosses with tropical breeds have been developed in the tropics and subtropics. The *Santa Gertrudis,* for example, is a cross between the Brahman and Shorthorn breeds. Dual-purpose sheep breeds such as the *Columbia* or *Corriedale* have been developed to produce high-grade mutton and good-quality wool under varied climatic conditions. Crossbreeding has also overcome, to some extent, the low lard-producing characteristics of hogs in hot climates.

INSECTS AND DISEASES

Many of the restrictions on the productivity and regional distribution of both plants and animals are pathological, that is, they result from the effects of insects or diseases. Climate, nevertheless, exerts an indirect influence, for insects and diseases have rather narrow climatic limits for maximum development. A favorable season for enemies of crops can result in a major catastrophe. Even short-term weather conditions may determine the extent of development and spread of a plague. That the relationships of insects and disease to climate are complex can be seen from the fact that weather conditions sometimes favor parasites or maladies which make inroads on insects. Moreover, the food supply of insects is controlled to varying degrees by weather and climate. Various types of plant enemies, such as mildew, rusts, scabs, and blights, reproduce and spread most rapidly under conditions of warmth and high humidity. Spores of fungus diseases are spread by the wind, making control very difficult. Wheat, barley, and certain legumes are largely excluded from humid climates because of disease problems rather than direct climatic factors. Coffee is confined to cooler uplands in the tropics primarily because of disease problems in the hot, wet lowlands. A host of mid-latitude crops are unsuited to the tropics for pathological reasons, even though the climate might appear to be directly favorable. A great many diseases and insects which attack plants are kept in check by seasonal fluctuations in temperature or moisture or both. For example, in the areas of the southeastern United States which are infested with the cotton-boll weevil, it has been estimated that 95 percent of hibernating adults die during the winter. Thus, the winter has a marked bearing upon the incidence of boll weevil damage in the summer. The incidence of diseases and insects harmful to plants is considerably less in high latitudes; this is due to the absence of host plants as well as to low temperatures. In other words, the same climatic factors that restrict diseases and insects often limit crop production as well. Wind influences the migration of insects, especially if the air is warm. The direction of migration is with the prevailing wind, and upper winds may be dominant in the distribution, for flights of insects sometimes reach heights above the surface winds.

The density of airborne insects is much greater near the surface when the air is relatively calm, but turbulence carries them upward. In North Africa and Southwest Asia locust swarms are associated with the intertropical convergence zone. Rain along the convergence zone provides moisture required for breeding as well as for vegetation upon which the larvae feed. The northerly and southerly winds tend to carry the swarms toward the areas of most favorable ecological conditions. (See Figure 12.15.)

Control of insects and diseases which damage plants belongs in the province of plant pathology rather than climatology. Any agricultural practice which provides an unfavorable environment for a crop pest has potential value for controlling it. Forced exposure to extremes of tem-

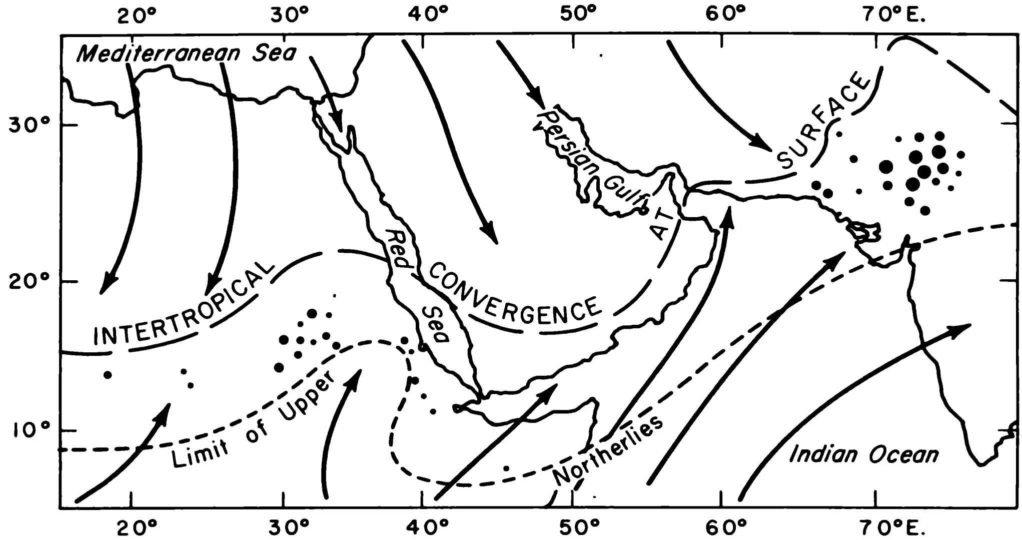


Figure 12.15
Locust swarms in relation to the intertropical convergence zone. Dot size is approximately proportional to number of swarms reported during a period of 20 days. (After Rainey and Bodenheimer)

perature, sunlight, or moisture conditions is one of the most practical approaches to the problem.

Insects and diseases which attack animals have a complicated relationship to climate and other environmental factors. Of the climatic factors, temperature and moisture are the most important, and as a result, there are broad correlations between infestation and both climatic regions and seasons. As in the case of crops, animals generally have fewer pathological enemies in the colder climates. Probably the largest class of malefactors in the livestock industries are internal parasites. Climate and daily weather have a marked influence on the distribution, rate of spread, and intensity of infestation of parasitic diseases. While developing in that part of the life cycle which is outside the animal, most parasites have limiting ranges of temperature and moisture requirements. Exposure to direct sunlight is usually fatal to eggs and larvae. Thus, control can be achieved to some extent by measures which enhance the destructive effects of climatic conditions on the eggs and larvae of parasites or on their intermediate hosts. Rainfall, wind, and sunlight render valuable assistance in maintaining sanitary conditions in pastures and feedlots. Their aid can be put to the best advantage by rotating animals from one enclosure to another.

Development of resistant types through selection and crossbreeding has met with some success in combating animal diseases. Certain breeds have natural immunities, and it is often possible to combine qualities of disease resistance and high productivity by crossbreeding. Among tropical breeds of domestic animals, there is commonly some degree of immunity to tropical diseases as well as heat tolerance, so that crossbreeding has improved both climatic adaptation and disease resistance of high-producing mid-latitude breeds in the tropics and subtropics.

It is difficult to separate the discussion of insects from that of diseases because, very often, an insect species is the carrier of a disease. Numerous insects, of course, attack animals directly. One can cite those which sting, chew, suck blood, lay eggs in the hair or skin, or merely swarm menacingly about the animal. Among the many insects that infect animals with disease, an outstanding culprit is the tsetse fly, which is limited to tropical Africa. There are several species of tsetse, none of which can endure low temperature nor extremely hot, dry weather. They live solely on blood, and, since they are disease vectors, they infect the animals on which they feed.

PLANT AND ANIMAL INTRODUCTION

The adaptation of crops and livestock to certain climatic optima is of great significance in connection with the introduction of a particular variety or breed into a new area. To the extent that climate influences the productiveness of crops or livestock, transfer of plants or animals to another region with a similar climate is likely to be successful. Tropical tree crops have been transplanted literally around the world. Coffee, believed to have been native to the highlands of eastern Africa, is now widely grown in tropical Latin America. The native rubber tree of Brazil has met with more favorable circumstances for commercial production in Southeast Asia than in its original habitat. Cacao, also indigenous to the Latin American tropics, is produced in the Guinea Coast region of Africa as well as in South and Central America. The introduction of such Mediterranean crops as citrus fruits, olives, and wine grapes into other dry summer subtropical regions is an example of migration of crops to similar climates. There is a huge list of crops which originated (or at least were acclimatized) in Europe, and which have been successfully carried into other mid-latitude climates by European emigrants. Nowhere is there a more striking example of wholesale transfer of plants and animals into a similar climate than the establishment of British agriculture in New Zealand. Breeding and selection play an important part in plant and animal introduction, as do economic and cultural factors. But the suitability of a climate for a particular crop or animal is a primary consideration which is not altered by the fact that many crops and animals have a wide range of adaptability to climate.

Whereas many crop plants have been introduced into new areas in the past largely on a trial-and-error basis, recent efforts have been directed toward detailed study of agroclimatic relationships to improve yields, quality, or disease resistance. As world agriculture has expanded and the demands for food have increased, plant introduction has come to mean more than the establishment of a new crop; it entails new varieties which have better characteristics for productivity in a particular environment. Varieties that have the desired characteristics for a given locality are sought in other areas where the climate is analogous. Regions having climates sufficiently alike to permit equivalent productivity of the same crop variety or crop combination are *agroclimatic analogs.* In the past simple temperature and rainfall data have dominated studies to determine analogs. As our knowledge of the relationships between plant physiology and climate improves, the analysis of regional climates for the purposes of plant introduction increasingly will use complex heat and water budget approaches as well as single factors such as radiation, light, or wind. In effect, the problem is one of applied climatic classification to achieve a specific objective.

Animals, being far more adaptable to variations in climate than plants, do not present so great a problem in connection with their introduction into a new region. Within broad limits climate influences animal introduction more through its effects on feed supplies than by direct means. When climates with sharply contrasting temperatures are involved, however, failure to acclimatize and to maintain satisfactory production levels may present a serious obstacle to livestock introduction. As stated earlier, crossbreeding and selection have overcome this problem to some extent.

Additional Readings

CHANG, JEN-HU, *Climate and Agriculture.* Chicago: Aldine Publishing Company, 1969.

———, "Potential Photosynthesis and Crop Productivity," *Ann. Assoc. Am. Geogrs.,* 60, 1 (1970), 92–101.

CRAWFORD, M., "Diseases of Livestock as Influenced by Weather," in *Biometeorology,* ed. S. W. Tromp. London: Pergamon Press, 1962, pp. 162–73.

FUQUAY, D. M., "Mountain Thunderstorms and Forest Fires," *Weatherwise,* 15, 4 (1962), 149–52.

HOGG, W. H., "Meteorology and Horticulture," *Weather,* 19, 8 (1964), 234–41.

JOHNSON, C. G., "The Aerial Migration of Insects," *Scient. Am.*, 209, 6 (1963), 132–38.

PENMAN, H. L., "Weather and Crops," *Quart. Journ. Roy. Met. Soc.*, 88, 377 (1962), 209–19.

REIFSNYDER, WILLIAM E., AND HOWARD W. LULL, *Radiant Energy in Relation to Forests*. Tech. Bull. 1344. Washington: U.S. Department of Agriculture, 1965.

RILEY, J. A., "Moisture and Cotton at Harvest Time in the Mississippi Delta," *Mon. Wea. Rev.*, 89, 9 (1961), 351–53.

SCHEIN, RICHARD D., "Biometeorology and Plant Diseases," *Bull. Am. Met. Soc.*, 44, 8 (1963), 499–504.

STANHILL, G., AND M. FUCHS, "The Climate of the Cotton Crop: Physical Characteristics and Microclimate Relationships," *Agri. Met.*, 5 (1968), 183–202.

TAYLOR, JAMES A., ed., *Weather Forecasting for Agriculture and Industry*. Cranbury, N.J.: Associated University Presses, Inc., 1972.

WANG, JEN-YU, *Agricultural Meteorology*. Milwaukee: Pacemaker Press, 1963.

WENT, FRITZ W., "Climate and Agriculture," *Scient. Am.*, 196, 6 (1957), 82–94.

chapter thirteen

Climate, Transportation, and Industry

Decisions involving weather and climate affect every economic activity in some way. This chapter treats the relationships between atmospheric conditions and representative problems in transportation, communications, manufacturing, and construction. Application of meteorology and climatology for improvement of efficiency in these fields of human enterprise entails decisions at both the planning and the operational stages. Whereas climate often is an important factor in long-range planning, as in the selection of a factory site, weather has a dominant role in daily operations. Wise decisions clearly depend on an understanding of the significant atmospheric processes as well as the requirements of a specific economic activity.

AVIATION

The acceleration of research in atmospheric science during the twentieth century has paralleled an increase in air travel, for weather information is vital to the safety and efficiency of aviation. As man ventures farther into, and through, the upper atmosphere knowledge of weather processes increases, and at the same time it becomes more necessary in planning and executing air and space travel. Since aviation is based on the earth's surface, weather conditions in the lower levels are as important as those in the upper air. Nor can climatology be neglected, for the probability of certain conditions enters into many decisions concerning both the functioning of air terminals and flight.

The application of climatology to aviation begins with the selection of an airport site. Assuming the need for an airport in a given area, the best site depends on a number of interrelated factors. Level land, absence of surrounding obstructions, and access to centers of population appear superficially to have little connection with climate. They may be factors affecting wind or visibility, however. Radiation fogs form more often in broad valleys, and if the site is to the leeward of industrial centers, smoke and other pollutants play a part in reduced visibility. Climatological studies make possible the selection of a site which experiences a minimum of weather hazards such as poor visibility, low clouds, turbulence, wind shifts, and thunderstorms. Commercial air terminals must be near cities but in locations that enjoy the maximum of acceptable visibility, high ceilings, and favorable winds. The ideal site is windward of air pollution sources, on a modest upland that provides air drainage into adjacent valleys while lying below mean prevailing cloud bases, and well removed from marshes, lakes, or other moisture sources. Prevailing winds determine the best orientation of runways to permit aircraft to take off and land into the wind. Average wind speeds as well as directions are taken into account; wind from some directions may be so light as to have little influence on traffic. Modern jet transports can land on dry runways in cross-wind speeds as great as 30 knots. Aircraft carriers align the runway with wind direction, and the speed of the ship becomes a factor in the relative wind speed along the flight deck. A flight cannot be judged by its speed and economy of operation in the air alone; low clouds, fog, or other weather conditions at takeoff or landing can affect the entire flight. If unfavorable weather is confined to a particular locality, landings may be possible at an alternate field; or the conditions may be temporary, allowing the aircraft to circle the airport while awaiting improvement. Development of so-called "blind landing" devices is gradually resolving the problems of low ceilings and poor visibility.

KEY TO AVIATION WEATHER REPORTS......

LOCATION IDENTIFIER AND TYPE OF REPORT*	SKY AND CEILING	VISIBILITY, WEATHER, AND OBSTRUCTION TO VISION	SEA-LEVEL PRESSURE	TEMPERATURE AND DEW POINT	WIND	ALTIMETER SETTING	RUNWAY VISUAL RANGE	CODED PIREPS
MKC	15⦶M25⊕	1R-K	132	/58/56	/1807	/993/	R04LVR20V40	/⊕55

Figure 13.1
Example of an aviation weather report as transmitted by teletype. (NOAA)

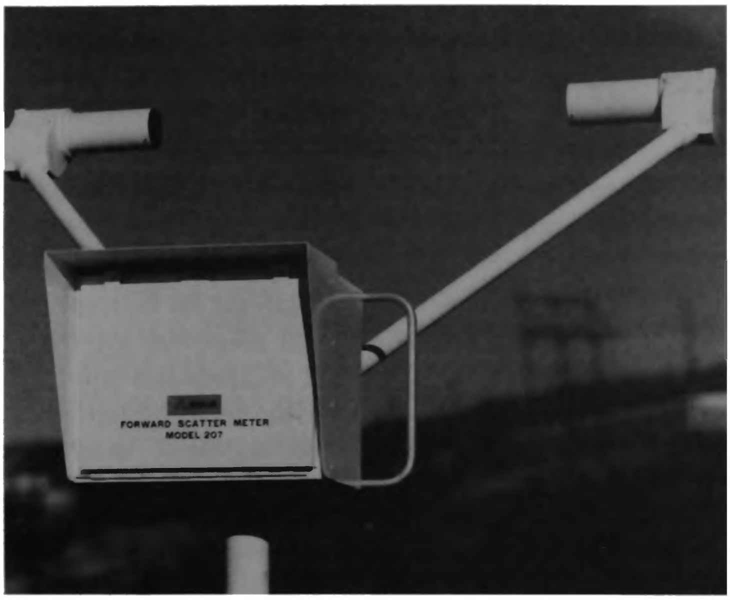


Figure 13.2 *Forward scatter meter. This instrument measures the reduction in transmitted light due to scattering by pollutants and supplements aircraft runway visibility observations.* (Photograph by Hutchins Photography, Inc., courtesy EG & G, Environmental Equipment Division)

To provide continuous information on airport weather the U.S. National Weather Service and the Federal Aviation Administration issue aviation weather reports at airports throughout the United States (see Figure 13.1). Comparable agencies serve aviation in other countries. Hourly reports are gathered and transmitted to other stations, principally by teletype, and to aircraft in flight by radio. They include data on cloud height and coverage, visibility, sea level barometric pressure, temperature, dew point, wind direction and speed, altimeter setting, precipitation, and special or severe weather conditions. Some airports are equipped to observe and report the runway visual range (see Figure 13.2). Special observations maintain the currency of data on sudden weather changes.

In the United States the Federal Aviation Administration prescribes minimum conditions for air safety. In addition to low ceilings and restricted visibility the pilot must, of course, make allowances for wind, icing, and other potentially dangerous phenomena. Most major airlines provide a special supplementary weather service for their pilots and dispatchers.

Knowledge of weather and climate enters into the design of aircraft and engines. Stresses likely to be encountered in severe storms, the possibility of damage by hail or lightning, and icing must be taken into account in structural engineering. Temperature variations, precipitation, and icing conditions affect power plant efficiency. Engines must be able to function at high elevations where pressure is much lower and where the temperatures may be 75C° or more below those at the point of takeoff. Aircraft flying from bases in winter or in high latitudes experience operational difficulties not found in the tropics. Special hangars or preheating devices are required to overcome certain problems arising from extremely low temperatures.

FLYING WEATHER

Contrary to popular belief, an arc of a great circle may not be the fastest nor the most economical route for air travel. Moreover, it may not be the safest nor the most comfortable route if unfavorable flying weather intervenes. Where possible a flight path should take advantage of tailwinds and avoid headwinds. Since wind speeds vary with altitude, this suggests a selection of the proper flight level as well as the best horizontal course. By climbing into the stream of strongest tailwinds an airplane can reap full benefit in savings on fuel and gains in ground speed. Except for very low flights, pressure distribution is a guide to route planning; gradient winds blow along isobars. Rarely, however, is wind direction exactly parallel to an air route; consequently an aerial navigator must know the true wind direction and speed in order to make adequate adjustments for drift. The extent to which it may be feasible to alter the flight path depends on the length of the flight, the locations of origin and destination with respect to atmospheric motion systems, and the possibility of encountering storms. It may be expedient to avoid the worst frontal weather in an extratropical cyclone rather than seek tailwinds. Unfortunately, air traffic density often makes it necessary to restrict aircraft to predetermined routes regardless of headwinds. In the case of long-range flight (for example, across entire continents or oceans) the route can be adjusted more easily to the pressure patterns, and therefore to the winds, at upper levels. The westerly jet streams have made it possible to achieve some remarkable records on west-to-east flights, even though an airplane may have deviated appreciably from the great circle route. Overseas flights often follow routes in winter different from those taken in summer in accordance with latitudinal shifts in wind and pressure systems.

For supersonic aircraft, winds are less significant on high-altitude flights of less than about 6,000 km. But the efficiency of supersonic jet engines is greater at lower air temperatures (and greater air densities), and it may be feasible to deviate from the great circle route in order to travel through colder air and thereby reduce fuel consumption.

Although powered aircraft operate most efficiently with tailwinds, gliders (or sailplanes) depend upon headwinds and vertical air currents to maintain altitude, and they have a limited range. Rising air currents over slopes, heated surfaces, or along zones of converging winds are favorable for lifting a glider and often are useful to light airplanes as well. Downdrafts, on the other hand, are to be avoided. Downslope winds along the margins of uplands create especially hazardous flying conditions. A pilot of an airplane or glider approaching a mountain may find that he is unable to maintain a safe altitude to fly above the mountain if the downslope flow is strong.

The use of the aneroid altimeter makes it necessary to keep abreast of changes in barometric pressure in flight. An altimeter set for an airport with relatively high pressure will indicate too high an elevation in an area with low pressure. (See Figure 13.3.) As a result, a pilot would tend to crash-land short of the runway or encounter obstacles he should otherwise clear easily. Conversely, an altimeter setting made in an area of low pressure indicates too low an altitude where the sea-level pressure is higher, and the possible result is a landing beyond the runway. Vertical air currents also affect aneroid altimeters, for they produce appreciable fluctuations in barometric pressure. A downdraft along a mountain slope is accompanied by higher pressure and therefore by an altimeter reading that is too high. This is another reason for extreme care when flying in the vicinity of mountains. An updraft not only tends to boost the airplane over the mountain but also produces an altimeter reading lower than the actual altitude, thus increasing the margin of safety. Where there is a choice it is better to fly on the windward than on the leeward of mountains. Other errors in altimeters result from temperature differences. In cold air the pressure decreases more rapidly with

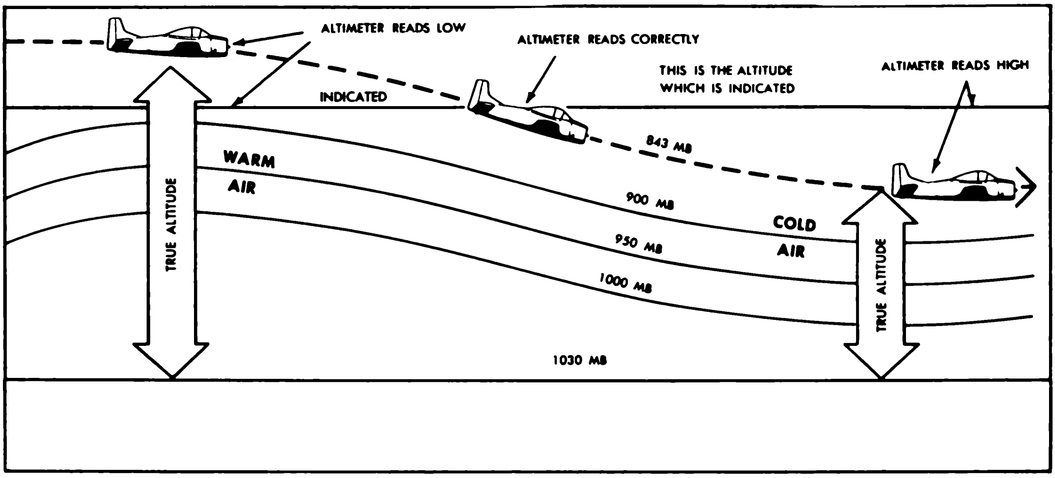


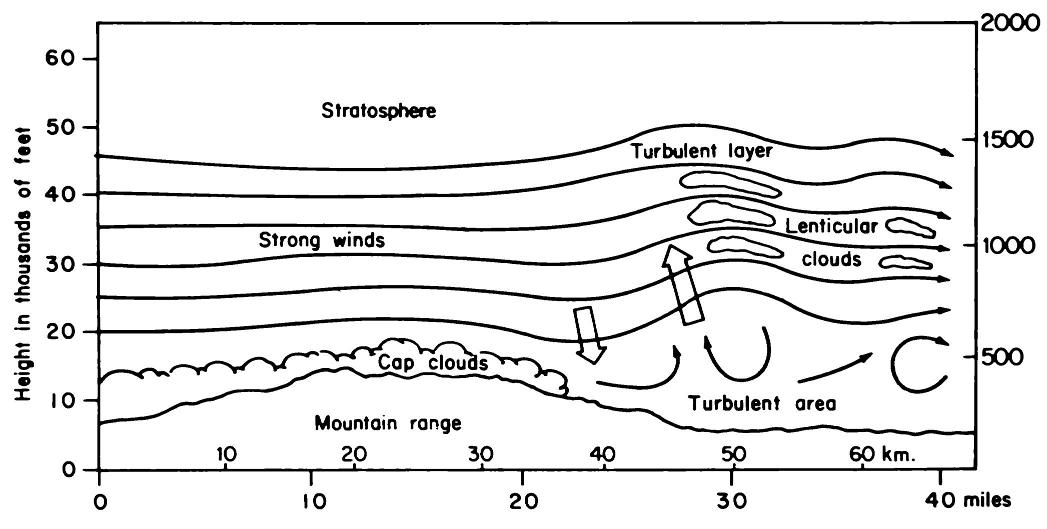
Figure 13.3
Effect of horizontal temperature and pressure differences on reading of aneroid type altimeter. (U.S. Air Force)

height, making the altimeter read too high; in warm air the pressure decrease is not so rapid and the altimeter of a plane in flight reads too low. Again there is danger in mountainous areas, where cold air may indirectly result in an altimeter reading higher than the true altitude. Altimeters which employ the principle of radar to indicate the elevation above the earth's surface are not subject to major errors due to pressure differences.

TURBULENCE

Atmospheric turbulence causes an uncomfortable ride and increases the strain on both aircraft and pilot, not to mention passengers. In its most violent forms it can be catastrophic. Four principal conditions lead to turbulence: thermal, frictional, frontal, and that associated with upper-level waves. Thermal turbulence is produced by vertical convection currents over a heated surface or by advection of cold air over a warm surface, and is accentuated in an air mass that already has an unstable lapse rate. Frictional (sometimes called mechanical) turbulence results from wind blowing over rough terrain. Its effects are often combined with convection currents when an irregular surface is heated or cooled unevenly. A strong flow of stable air across a mountain range frequently develops wave action with updrafts and downdrafts that extend to several times the height of the mountains and for a distance of 150 km or more downwind. Turbulence in such a *mountain wave* is usually greatest at low levels and again near the tropopause to the lee of the mountains. (See Figure 13.4).

Figure 13.4
Mountain wave turbulence in the lee of a mountain range.



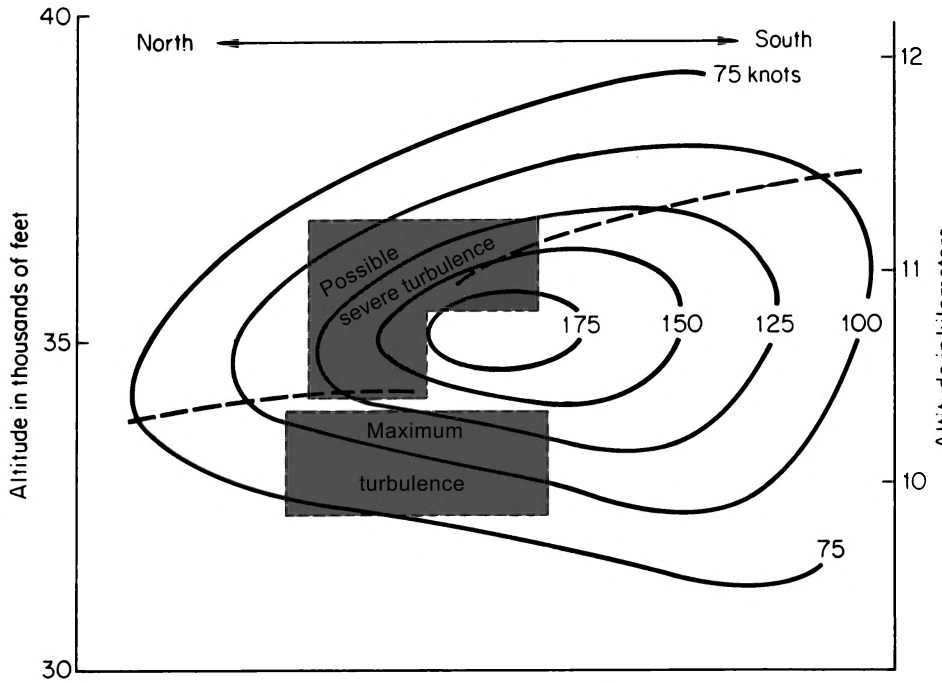


Figure 13.5
Wave-induced turbulence in the vicinity of the polar front jet stream. Solid lines are isotachs in the jet-stream cross section.

Along a cold front the wind shift and the lifting of warm air above the cold may cause wind shear, that is, pronounced differences in wind velocity in adjacent air streams. Aircraft sometimes encounter layers of turbulent wind shear when climbing or descending through a temperature inversion at any altitude up to the tropopause. Warm air moving over a ground inversion layer, as in a valley at night, creates a thin zone of wind shear that can be especially dangerous. Although turbulent conditions may have accompanying clouds as warning signs, they also disturb clear air. Wave-induced turbulence (*WIT*) is common in the vicinity of a jet stream, where both horizontal and vertical shear are generated (see Figure 13.5).

AIRCRAFT ICING

Ice formation on aircraft is one of the great hazards to flight in the troposphere. Icing may result when liquid water droplets are present and the temperature of the exposed aircraft surface is at or below freezing. It increases with air speed up to about 400 knots and then decreases because of heat produced by friction along the fuselage and projections. Clouds containing supercooled droplets are breeding places for aircraft icing. In turbulent air the drops are large, and when they strike the cold surface of an airplane they spread and freeze to form *clear ice.* In smoother air the small droplets freeze immediately upon striking the leading edges and take the granular or crystalline form of *rime ice*. Often the two types occur together, but clear ice is more tenacious than rime and is prevalent in turbulent clouds. It forms most commonly at temperatures between −25° and 0°C and can result from the fall of rain through freezing air. It creates a blunt edge on wings and other leading edges and thus interferes with their aerodynamic properties. Clear ice also adds dangerous weight to an airplane. Snow or ice pellets fall from clouds in which icing can be expected. Freezing rain, on the other hand, falls from warmer levels and can be avoided by climbing into the warmer air. Rime icing is most frequent in stratiform clouds at temperatures of −20° to −10°C and tends to accumulate in the form of sharp projections from leading edges. It does not disrupt air flow as seriously as clear ice but is, nevertheless, cause for caution.

In turbojet engines ice may form on guide vanes or other structures at the air intake, or in the fuel system, and lead to reduced engine thrust and eventual turbine failure. In piston engines carburetor icing can occur in clear air at temperatures above freezing if the relative humidity is high, but it also forms in clouds. The expansion of air and vaporization of gasoline at the fuel intake causes a decrease in temperature and condensation of water vapor. If the temperature drops below 0°C, the moisture is deposited as frost in the carburetor, and can reduce power or possibly cause engine failure. Carburetor icing is most common at air temperatures of −10° to 25°C in moist air.

FLYING IN STORMS

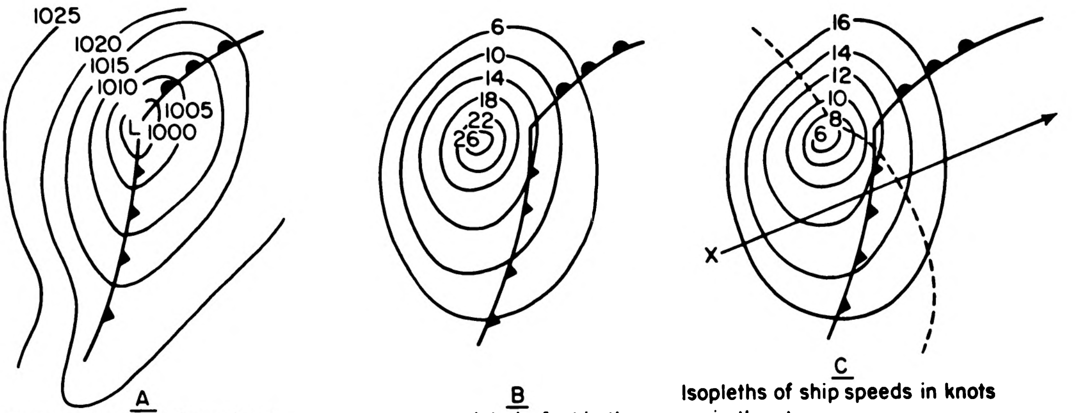
Flight through storms entails all the hazards of wind, turbulence, precipitation, low temperatures, poor visibility, and icing. Entry into a tornado is out of the question. Hurricanes are certainly not for routine air travel, but they have been entered for carefully controlled experiments and observations. Next in order of violence are thunderstorms, which because of severe turbulence, icing, and possible damage by lightning or hail are best avoided by flying above, below, or around them, depending upon circumstances. In a mid-latitude cyclone a variety of weather conditions may affect flight. Low ceilings and poor visibility accompany the typical warm front, and there may be hidden thunderstorms. Icing is likely in the cool upper levels, and in winter, sleet or freezing rain are hazards. Air mass weather in the warm sector of a cyclone may include fair weather cumulus, daytime convective turbulence, or thunderstorms. The cold front is usually the most dangerous portion of the mid-latitude cyclone. Frontal thunderstorms and a sudden change in wind direction and temperature are common features which bear upon safety and navigation. Occluded fronts tend to combine the flight conditions of the warm and cold fronts. Both warm- and cold-front occlusions have warm-front weather in advance of the occlusion, and both have upper-front weather like that of the cold front. (See Chapter 5, Occluded Fronts.)

In planning a flight the pilot has the aid of weather maps, upper-air data, and forecasts. His route and altitude are determined in relation to (and with respect for) weather conditions along the way. In flight, he is aided by radio reports and possibly by radar and weather instruments in the aircraft, but he must understand basic meteorological principles so that he can analyze unexpected conditions and alter plans accordingly to assure maximum safety. In regions where there are few weather reports, he must be both an observer and a forecaster and the success of his flight may well depend on his ability to interpret observed weather phenomena in the light of the climatology of storms.

WATER TRANSPORT

Mariners have a long history of weather problems. Sailing ships depended on favorable winds, and history records many instances where wind was the crucial factor in naval engagements between fleets powered by sails. Strong gales probably accounted for several early voyages of "discovery" as well as for many lost ships. The discovery of America by Columbus was influenced, in part, by the trade winds. The pattern of prevailing winds and pressure centers over the ocean determined, to a large extent, the major trade routes of sailing days. Even today, when steam and motor vessels dominate ocean shipping, wind direction and force are important in the safe and economic operation of ships. Navigation is especially hazardous in coastal waters when strong winds are blowing. On the open sea, a favorable wind can give increased speed to a ship, whereas a headwind slows the ship and increases fuel consumption. The pressure of wind against a ship varies as the square of the wind speed. A doubling of the speed increases wind pressure four times. Ships moving against headwinds must fight the relative wind speeds which represent the true speed plus the speed of the ship. Wind from the stern increases a ship's speed by about 1 percent, but headwinds decrease the speed by from 3 to 13 percent, depending on the size of the vessel and its load. A longer route, planned in relation to winds and storms, can result in greater economy of operation, less time at sea, and a smoother trip.

In rough seas it is difficult to hold a course and sometimes is necessary to change the course in order to reduce the effect of waves and swell.



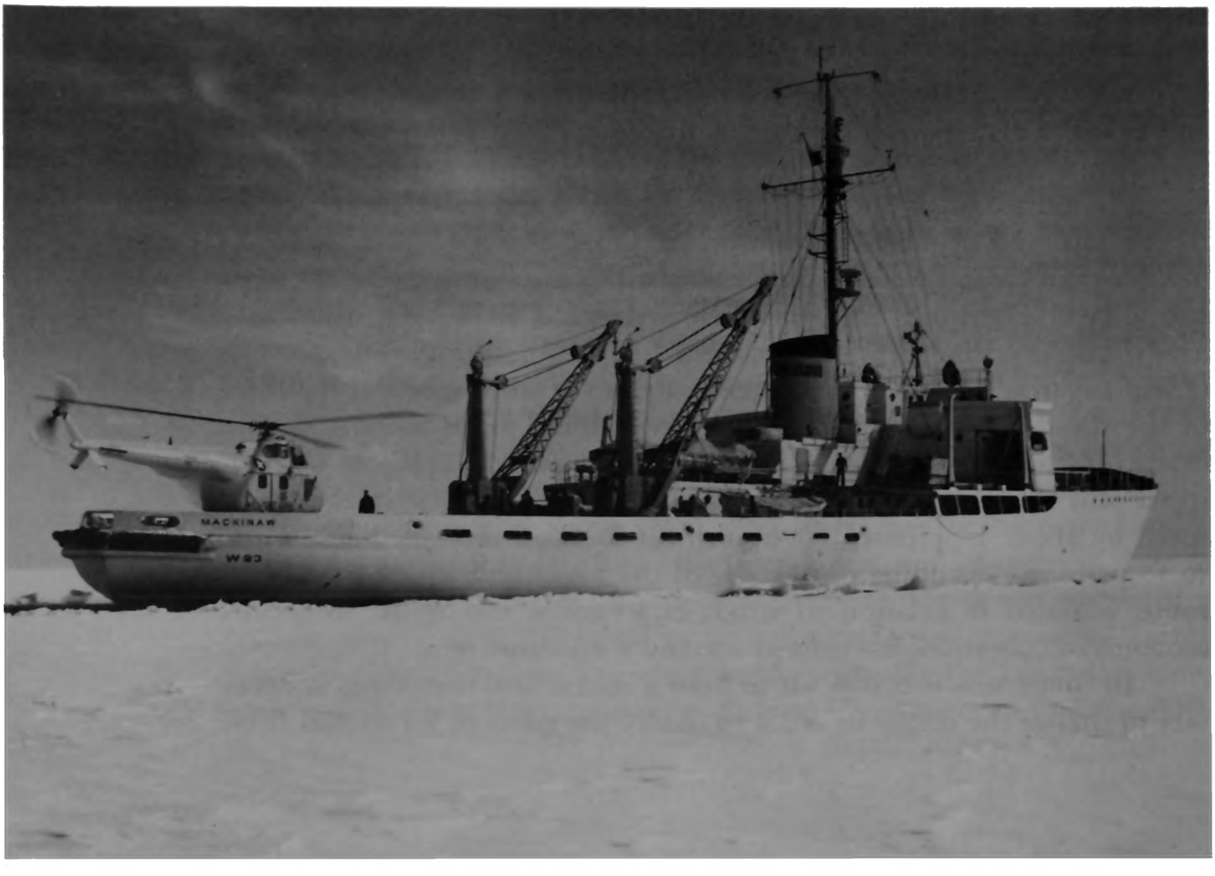
A
Sea surface weather chart

B
Wave height in feet in the area of A

C
Isopleths of ship speeds in knots in the storm area
Dashed line indicates relative distance attained in 24 hours by a ship moving northeastward from point X

Figure 13.6
Relation of ship speed to wave height in the area of an extra-tropical cyclone. (After J. J. Schule, "Weather Routing of Ships," in *Meteorology as Applied to Navigation of Ships,* Technical Note No. 23, Geneva, WMO, 1958)

Figure 13.7
Coast Guard icebreaker Mackinaw *opening a passage in the Great Lakes.* (U.S. Coast Guard Official Photo)



(See Figure 13.6.) Harbors in sheltered coastal locations, such as bays and river mouths, have always been desirable because of the relative absence of high winds and rough water, but even in the snuggest harbor damage can occur to vessels which are not properly moored. High or gusty winds make a landing dangerous. In cold climates, wind often piles up ice in harbors and increases the possibility of damage due to collision with ice chunks.

Since the advent of the steamship, sailing has become largely a sport, for the most part confined to coastal and inland waters. But the importance of wind to sailing craft, both in attaining speed and in relation to safety, has not diminished. Erratic winds on mountain lakes or along rugged coasts make extreme care mandatory in operating small pleasure craft. Many pleasure-boat owners use their craft only a few days each year and are unfamiliar with the principles of safe navigation as well as with local weather hazards. Sudden squalls have, all too frequently, brought disaster to small boats caught too far from shelter.

Next to wind, visibility is the most significant weather element in water transportation. Fog is the most common obstruction to visibility over water, although industrial smoke may intensify it in harbors. In the open sea, fog interferes with the accurate determination of the ship's position and increases the danger of hitting another ship, rocks, or icebergs. Along coasts where upwelling of cold water occurs, fog is common, and particular care must be exercised to avoid shallows or other danger points. Beacon lights afford some aid as warning devices, but foghorns have a greater range in dense fog. Modern foghorns commence operation when fog becomes dense enough to interrupt a beam of light between two fixed points. Radio, radar, and electronic devices for sounding ocean depths have considerably lessened the danger of collision or running aground in fog. Even so, dense fogs can bring harbor traffic to a standstill and many marine accidents due to fog occur every year.

Low temperatures affect loading and unloading activities and must be met with extra protection for cargo and passengers. If the temperature is low enough to form ice, otherwise navigable waters may become impassable (see Figure 13.7). In arctic seas, ocean traffic is suspended for several months each year. At lower latitudes, icebreakers are employed to keep navigable channels open. On the Great Lakes, ice builds out from the shorelines and thus disrupts shipping, even though the lakes are rarely frozen over their entire area. The ice is generally thickest in protected bays; straits become congested with ice carried in by the wind and currents. Broken ice, or "slush," can damage steering gear and propellers and is not effectively moved by an icebreaker, for it tends to close around the ship. Heavy shipping moves on the Great Lakes for only eight or nine months a year, so that it is necessary to stockpile materials such as coal

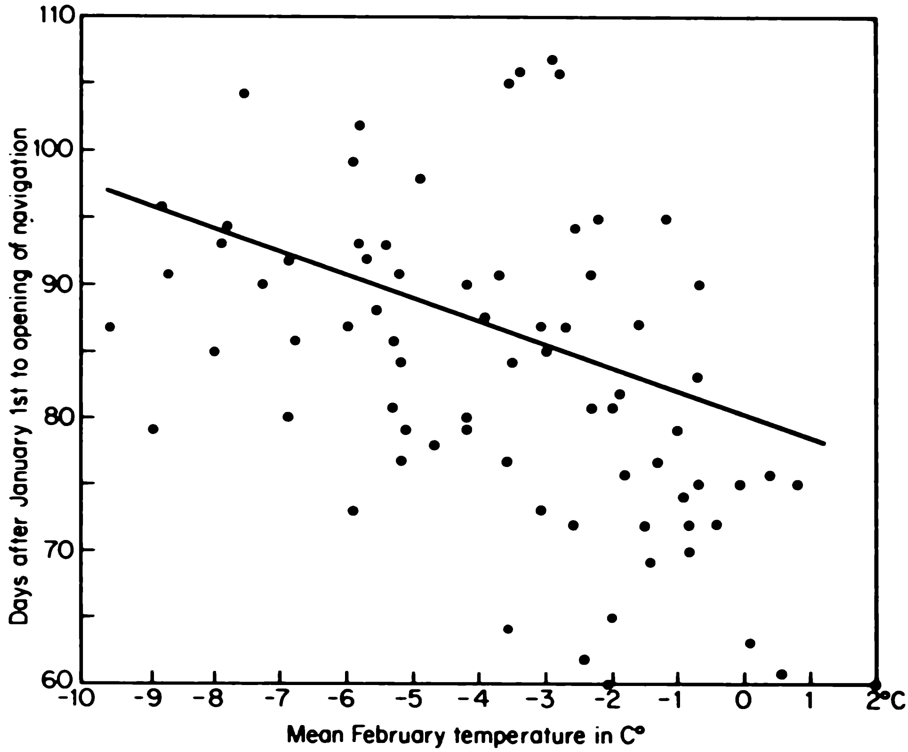


Figure 13.8
Opening dates of navigation at Detroit, Michigan, and mean temperatures of the preceding February. (Data from NOAA)

and iron ore in the navigation season to carry industry through the winter. Dates of opening of navigation on the lakes correlate broadly with the mean temperatures of the preceding February. In general, low February mean temperatures are followed by late clearing of ice from harbors. (See Figure 13.8.) Forecasts based on these relationships enable shippers to prepare their equipment with the minimum of delay and expense before seasonal operations begin.

Similar problems relating to ice on rivers, canals, lakes, and harbors occur throughout the higher middle latitudes. The net effect is to produce a seasonal rhythm in waterborne traffic with variations in the length of the shipping period from year to year and place to place.

Precipitation is ordinarily not a major factor affecting shipping. Loading and unloading are disrupted if perishable cargo is involved, and special protective measures are necessary for loaded cargo. Heavy rain or snow may affect visibility. Storms incorporate several of the adverse weather elements, and, depending on their severity, it is sometimes desirable to avoid them. Hurricanes, in particular, call for "evasive action." Wind speeds and sea swell are greatest at the front of a hurricane, and make travel to the rear of the storm advisable. Where forecasts are available, it is often worthwhile to alter a ship's speed in order to arrive in port before or after a storm reaches the port. Since the danger of damage resulting from severe storms is greater near the coast or in poorly protected harbors, large ships usually head for the open sea when a storm with high winds is expected.

Radio weather reports and marine forecasts are broadcast from many coastal stations as an aid to ocean navigation. Much of the interpretation of these reports, however, is left to the mariner; indeed, valuable weather data upon which marine forecasts are based are contributed by ships at sea. Along the Atlantic Coast of the United States and Canada, reports from land stations to the west are fairly abundant, and they can be used to construct a weather map and make a forecast which is far more reliable than guesses based on weather signs. Off the Pacific Coast, weather reports from continental land stations are of limited value while a ship is at sea because the paths of the storms are generally from west to east. Between the Aleutians and the Hawaiian Islands there are no Pacific islands. It is therefore necessary to rely on sparse reports from weather ships, commercial vessels, and aircraft to analyze weather trends on the basis of observations.

RAILWAYS AND HIGHWAYS

Railroads would seem to be little affected by weather because of their permanent all-weather track systems. In fact, trains are likely to be on the move in storms which have brought automobile traffic to a halt and grounded all airplanes. But railroads are far from immune to weather hazards, and their operation must be constantly geared to weather changes. Severe storms may damage tracks, bridges, signals, and communication lines. Floods, heavy snowfall, and earth slides are related menaces. Heavy snow and avalanches are particularly troublesome in mountains, and it has been expedient in many cases to build expensive tunnels to avoid high-altitude weather as well as steep grades. Low visibility due to fog or precipitation calls for extra caution and decreased speeds. Passenger schedules are disrupted not only by the direct effects of bad weather on tracks and equipment but because many people forsake the airlines or highways for the train and extra time is consequently required for station stops. Passenger traffic to beaches, resorts, ski areas, and other holiday locations is closely related to weather conditions. Seasonal weather greatly affects tourist travel; the general movement in the United States is southward in winter and northward and westward in summer.

In the movement of freight on railroads, special attention must be

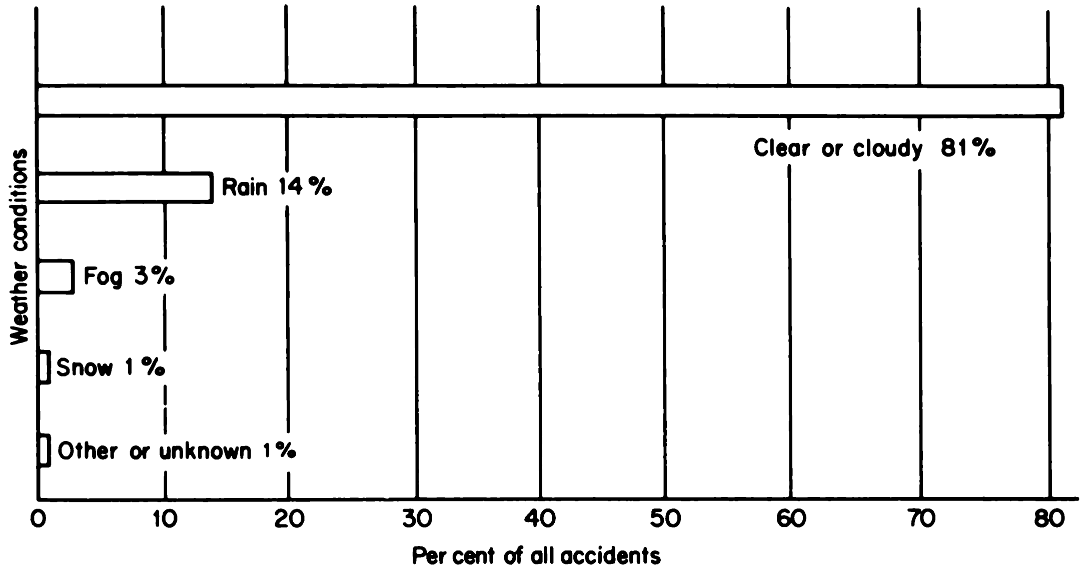


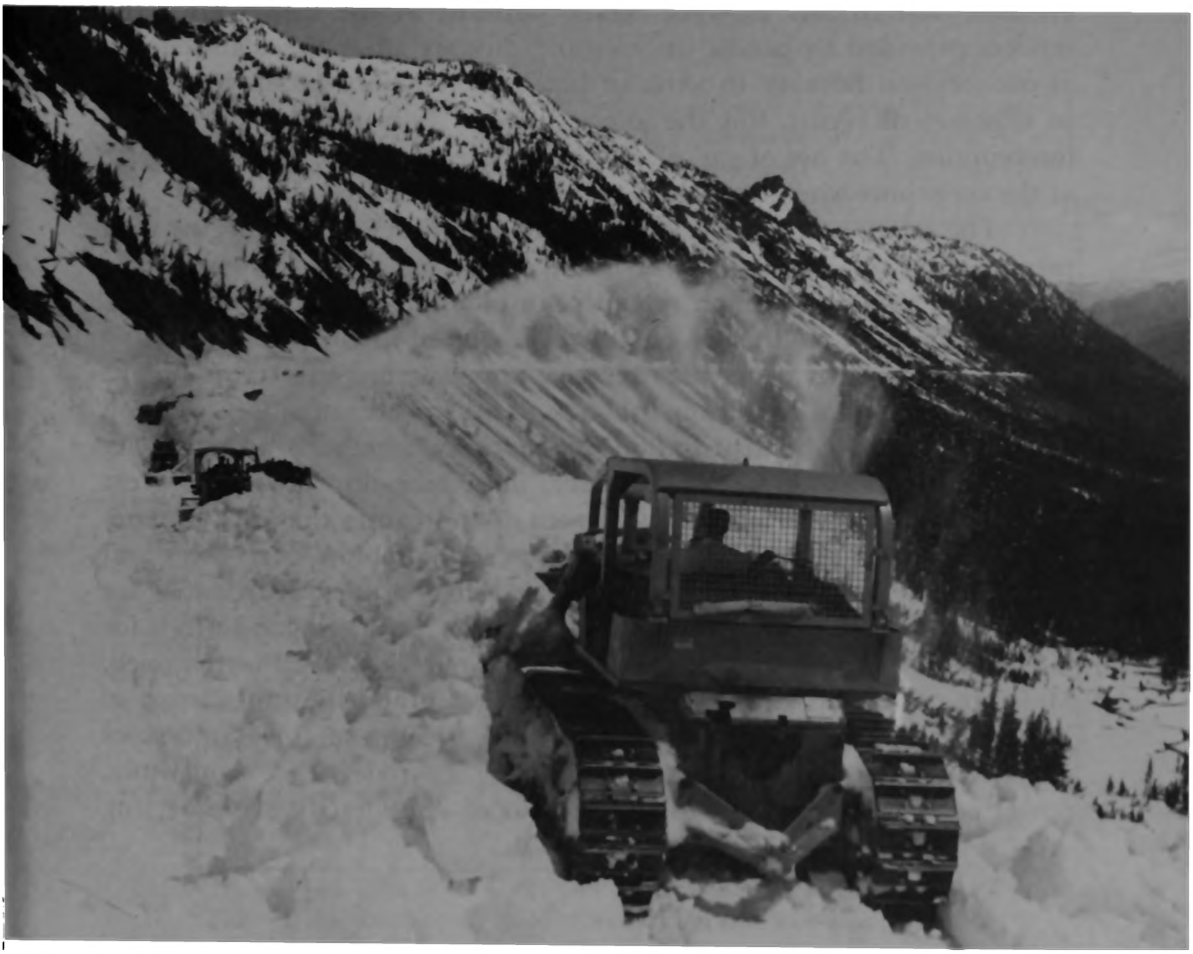
Figure 13.9
Percentage of motor vehicle accidents in different weather conditions. Based on total accident reports for 30 states. (Data from National Safety Council)

given to perishable goods and livestock to prevent losses during unfavorable weather. There are seasonal variations in the types and quality of freight. Fresh fruits and vegetables must be moved immediately after harvest, the time of which is determined in part by weather. Freight lines serving ports which are icebound in winter necessarily experience much of the seasonal variation in traffic imposed on water transport. Perishable freight for export must be routed to other ports; nonperishable goods may be stockpiled.

The major consideration in the relation of weather to the use of highways is safety. Poor visibility, slippery surfaces, and gusty winds increase the hazards of driving. Snow, freezing rain, and fog create especially hazardous conditions calling for mechanical equipment in good working order and for caution on the part of drivers. Good weather, especially on weekends, brings large numbers of pleasure drivers onto the highways and traffic problems and accidents increase. Indeed, most accidents occur in relatively good weather. (See Figure 13.9.) Highway patrols experience fluctuations in activity which are closely correlated with fine holiday weather as well as bad weather. It does not follow that the alarming increase in traffic deaths is caused by the weather, however. Weather is a factor only insofar as it may create more hazardous conditions and a need for greater driving care.

Many of the maintenance problems on highways and roads result from weather and climate. The expense of erecting signs to warn motorists of hazards associated with weather is great. Unsurfaced roads suffer heavy damage from precipitation, and, when wet, are damaged by traffic. Even paved highways undergo erosion at the shoulders, which may lead to collapse of the roadbed. Floods can wash out bridges, underpasses, and even entire sections of road, or they may carry debris onto the highway. Gales blow trees, utility poles, and other obstructions onto the roadway. Alternate freezing and thawing causes frost-heave in the road surface and subsequent breakup of the roadbed under heavy traffic. It is often necessary to prohibit heavy vehicles from wet or thawing roads to reduce surface damage. Applications of sand increase traction, and salt speeds melting when roads are icy. Bridges tend to develop icy surfaces sooner than roadways because they are exposed on all sides to cold air. Snow is one of the most expensive highway maintenance problems in the

Figure 13.10
Snow removal on Chinook Pass, Cascade Mountains, Washington. (Washington State Department of Highways)



middle and high latitudes and in mountains. (See Figure 13.10.) Where drifting is common, snow fences or windbreaks along the windward sides of highways help induce drifts away from the road. Tunnels and snow sheds are built in mountain passes to avoid deep snow as well as avalanches. Because of the excessive cost of keeping them open with snowplows, some passes are closed for all or a part of the winter. Snow removal crews and equipment must be ready for immediate action in snow areas; great sums are expended in order to keep roads open for traffic.

The need to keep city streets clear of ice and snow is even more urgent than on the highways. Sewers and gutters easily become clogged by heavy accumulations of snow. Besides being a traffic hazard, snowy streets are an inconvenience to pedestrians. Snow of great depth cannot be merely plowed to the side but must be hauled away. If the temperature is not too low, it may be removed by melting with water. Installation of hot-water or steam pipes under streets is a possible solution to the snow problem.

PUBLIC UTILITIES

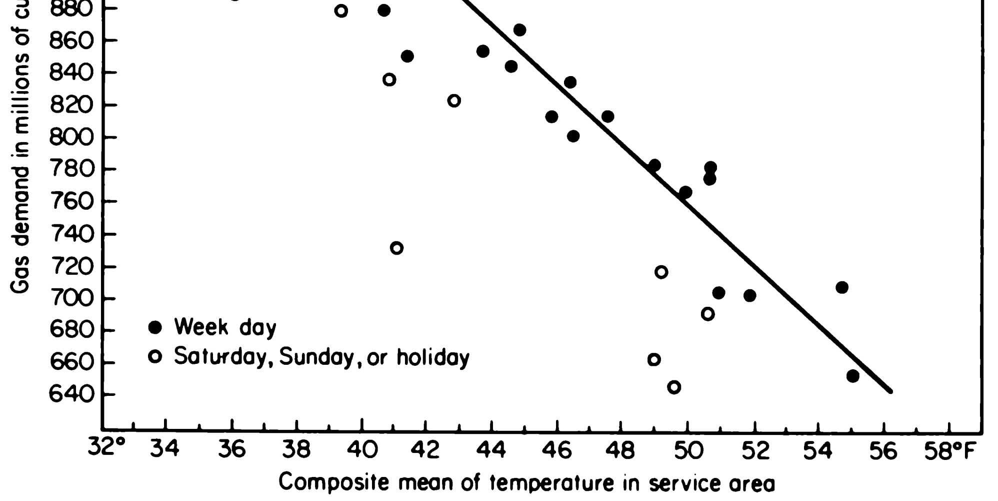
Weather and climate produce erratic patterns in the demand for the services provided by public utilities and directly affect the maintenance of the services. Damage to wires or pipelines not only entails losses due to expenses of repair, but the affected utility loses revenue during the interruption. The use of gas, electricity, or the telephone commonly rises at the very times when maintenance problems are at a maximum.

The relation of weather and climate to hydroelectricity begins with precipitation on a watershed. It is important for a power company to know what volume of runoff to expect in planning power commitments. The influences of weather on runoff are discussed in Chapter 11. During dry summer weather it may be necessary to release more reservoir storage to meet the needs of irrigation, fisheries, or other water uses. Construction of power lines for the distribution of electricity, whether that electricity is developed by steam or water power, must take into account a great number of climatic effects on equipment. Wind causes the greatest damage to power poles and lines; strong, gusty winds blow over poles and snap lines, or blow trees and other debris into the lines. Hurricanes are by far the worst storms for overhead power lines, but even good winds for kite-flying are bad for power lines, if kites are not kept well in the open. Moisture conditions affect electrical transmission and the functioning of insulators. Wet snow or freezing rain adds weight to lines and poles and makes them more vulnerable to wind. To remove ice from lines, power companies employ various methods of temporarily disconnecting a line from the main circuit and heating it with an abnormally high power load. Temperature fluctuations influence the operation of switches, transformers, and other equipment. In hot weather, lines expand and sag and are more easily damaged by high winds. Thunderstorms bring the dangers not only of wind and precipitation but also of lightning, which causes at least a temporary power failure if it strikes a line. Weather forecasting aids in anticipating heavy power demands and extra maintenance problems. As a storm or cold wave moves across the country, the load is often sustained by rerouting electricity around affected areas or by bringing emergency generators into use at a moment's notice. In view of the many problems which weather imposes on overhead lines, it may well be asked if it wouldn't be wise to put the lines underground. From a meteorological point of view it would, but the cost would be exorbitant.

The importance of weather information to gas companies is primarily in connection with consumer demand, although, as with other utilities, the maintenance of distribution facilities is rendered more difficult by severe weather. Delivery of natural gas over long distances cannot ordinarily be increased as rapidly as can electrical service, so that it is even more important to be able to anticipate future needs in the

Figure 13.11

Relation of demand for gas to air temperature in a service area. Data are for a single winter month. The solid line represents estimated demand for gas at different mean daily temperatures, averaged as a composite for the entire service area. (Data from Pacific Gas and Electric Company)



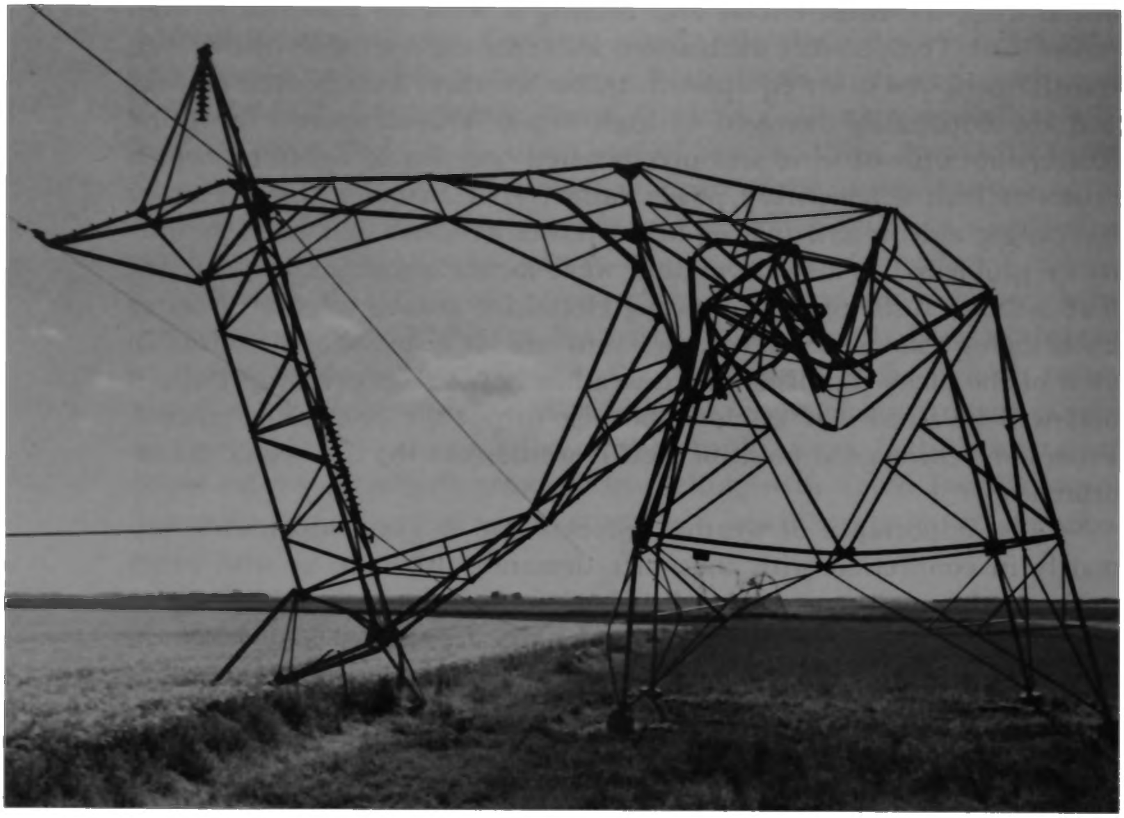


Figure 13.12
Electric power transmission towers damaged by wind. (Bureau of Reclamation, U.S. Department of the Interior)

light of weather forecasts. Temperature, wind, cloudiness, and precipitation all affect the use of gas. Of these, the most important is temperature. (See Figure 13.11.) The demand for more gas resulting from a cold wave must be met by dispatching gas from a remote source *before* the low temperatures occur. For the most efficient and economical service the gas dispatcher constantly refers to the meteorological conditions and forecasts for the area being served. Several gas companies have found it profitable to employ meteorologists who can interpret weather conditions specifically in terms of the problems of gas distribution.

Telephone and telegraph services have many of the same problems as power companies in maintaining their lines in all types of weather. (See Figure 13.12.) Since telephone and telegraph lines are closely linked throughout the United States, it is often possible to channel communications around affected areas in times of emergency. In bad weather, especially during major disasters, the public makes heavy demands on communications systems, and these media must be geared to meet maximum traffic and poorest operating conditions at the same time. Microwave transmission eliminates some of the line problems but introduces new difficulties because transmission is affected by temperature and moisture conditions in the atmosphere.

Radio communication and television are freed, to a large extent, from the problems of line maintenance, although antennas are subject to the elements. Indirectly, radio transmission and reception are affected by power failures due to storms, and where communication is vital emergency power sources must be provided. Directly, the transmission of radio waves is influenced by atmospheric conditions. Lightning, other electrical disturbances, and displays of auroras are usually accompanied by poor radio reception. Changing temperature and humidity characteristics of the air alter the refraction of radio waves and thereby influence their propagation. Temperature inversions and steep humidity gradients near the earth's surface tend to produce an excessive downward refraction of radio waves, thus limiting the distance to which signals reach. On the other hand, long-distance radio reception would not be possible were it not for the downward refraction of radio waves by certain of the ionized layers in the outer atmosphere.

Municipal and industrial water supplies are closely related to precipitation and the hydrologic cycle. Year-to-year and seasonal fluctuations in the amount of water available result from variations in precipitation. Thus, a city water system must be planned for dry years as well as average years to provide basic minimum service. Where the ground freezes each winter, water pipes must be placed well below the expected frost depth. In some instances, in regions with permafrost, water mains, sewer lines, and hot-water pipes have been placed in the same insulated tunnel to overcome the threat of freezing.

Sewage systems depend directly upon the water supply and therefore are subject to all its problems, whether caused by weather or not. Temperature has minor effects on sewage disposal; the speed of chemical and biological activity is increased under high temperatures and decreases in cold weather. Storm sewers are ordinarily separate from sanitary sewers to minimize treatment and disposal problems. Their proper design and construction depend upon knowledge of rainfall extremes and potential runoff.

MANUFACTURING

The influences of weather and climate on manufacturing are so many and varied that they defy a complete listing, let alone a thorough analysis, between the covers of one volume. They may be divided into two general categories: those influencing location and those affecting operations once an industry has been established. Climatic factors influence the location of a factory because of their effects on transportation, raw materials, labor supply, housing costs, manufacturing processes, and waste disposal. The relative importance of climatic factors depends on the type of manufacturing in question, and climate is frequently overshadowed by more pertinent economic considerations. Food-processing plants are likely to be located where the agricultural produce for manufacture is available. We can say that the availability of raw material is the primary factor, but certainly climate plays a fundamental role. Similarly, pulp and paper mills are located where there are abundant water supplies, which again presumes favorable climatic conditions. A factory demanding year-round access to water transportation would hardly be located on a harbor that freezes over for part of the year. The attractiveness of a climate to a large labor force may be a factor in inducing labor to move to a new area; at least it will be considered in the category of "fringe benefits." The cost of heating, or air conditioning, to provide satisfactory conditions for work and for certain manufacturing processes bears a direct relationship to climate. Factory buildings, warehouses, and other structures should be designed with climate in mind. Factories which dispose of wastes through stacks must be so located as to minimize air pollution over settled areas.

Weather conditions are reflected in almost every phase of operation of manufacturing plants. Storms cause workers to arrive late for work, hamper essential outdoor activities, cause damage to goods and equipment, or interrupt power. Some processes, notably in the chemical and related industries, have a narrow range of temperature requirements for efficient production. Wind and low relative humidity increase the fire hazard at manufacturing plants, especially if combustible or explosive materials are being processed or stored.

ATMOSPHERIC POLLUTION

Probably the most widespread problem confronting industrial meteorologists is that of atmospheric pollution around cities and factories, but it is not the problem of meteorologists alone. Weather and climate are not the causes of air pollution, but atmospheric conditions do greatly affect the rate of diffusion of contaminating agents, both horizontally and vertically. Since air pollution is far more common around cities than elsewhere and has increased with their growth, it is evident that it is a man-caused problem. Reduced vsibility due to pollution is related to the greater urban activity on weekdays. One can often locate a distant city by its cap of smoke and haze. There are always some foreign materials in the air—in liquid, solid, and gaseous forms. The increase in pollution has come from a multitude of sources. Smoke, dust, gases, and vapors from industrial operations account for a large share; heating systems, incinerators, motor vehicle exhausts, and evaporation of volatile liquids such as gasoline are other sources.

The meterological effects of pollution are largely concerned with visibility and solar radiation. Since many contaminants are hygroscopic they act as nuclei of condensation and create a haze or fog, further reducing visibility. Moreover, it appears that the water droplets thus formed are more stable than normal cloud droplets and do not evaporate so readily upon being heated. Oily substances, especially, tend to form a protective coating around a droplet, making it difficult to disperse. Originally the term *smog* denoted a combination of fog and pollutants. Solar radiation is appreciably reduced by polluted air in the daytime, and outgoing radiation is reduced at night. The net effect is a lowered diurnal range of temperature. Parts of the solar spectrum are transmitted selectively by different gaseous constituents in the air so that the quality as well as the quantity of insolation is impaired. Chemical reactions among various types of contaminants in the air produce new compounds that in some cases are more damaging than the original wastes. Certain of these reactions are photochemical, that is, they take place under the effects of sunlight.

Stable air and light winds or calms are conducive to concentration of pollutants at or near the source of contamination. For numerous reasons industries commonly locate in valleys and depressions for which stable air has an affinity and where temperature inversions are common. Temperature inversions are particularly suited to the formation of palls of smoke and industrial haze. As the warm fumes, gases, or airborne solids rise they cool adiabatically as well as by radiation and mixing. Since the air is warmer overhead in an inversion, they are soon at a temperature equal to that of the surrounding air. Therefore, they do not rise further. Cooling by radiation at night from the top of a smoke layer intensifies the effect, and the concentration of pollutants increases at and below the top of the inversion. Stable air, especially with an inverted lapse rate, often is accompanied by radiation fog, which combines with the pollutants to form smog. As already indicated, hygroscopic particles may hasten the condensation process. If the inversion layer is well above the surface, a pall will form with less drastic effects, but it will nevertheless inhibit the passage of radiation and may eventually "build down" to the surface. The most intense smogs usually develop when the top of the inversion is within 500 m of the ground. Pollutants tend to concentrate near the ground in a nocturnal surface inversion. By day, heating of the ground may create a layer of turbulent mixing which does

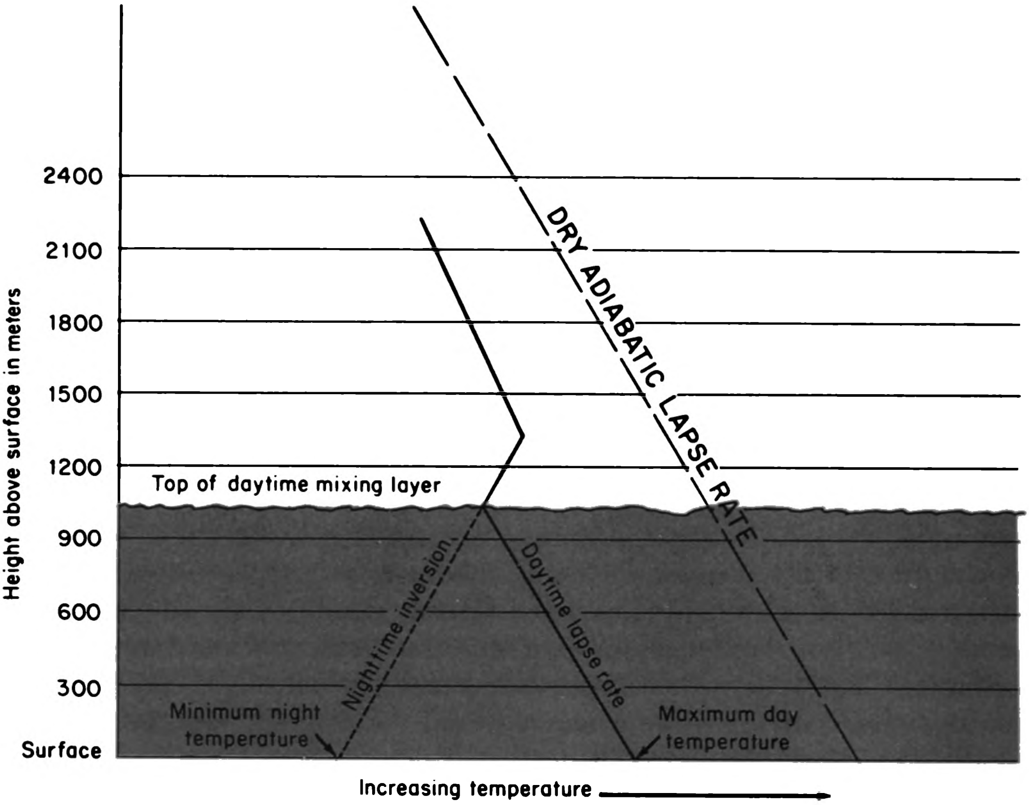


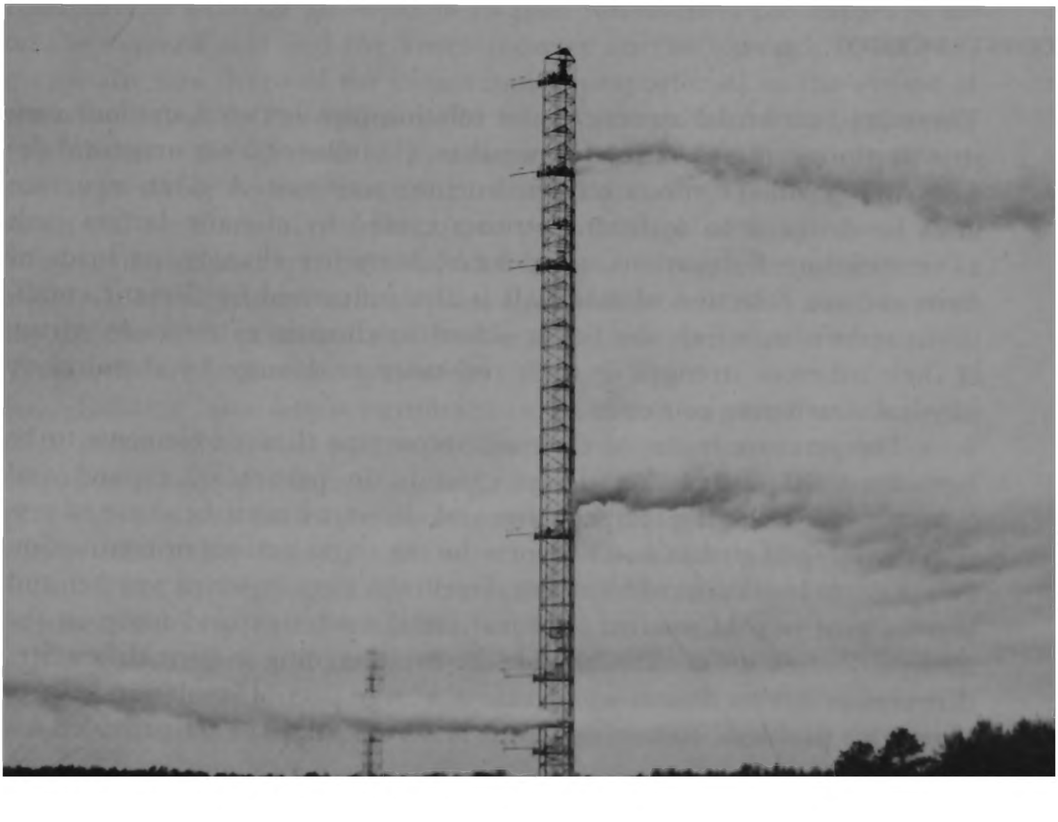
Figure 13.13
Height of daytime surface-mixing layer in relation to temperature lapse rate.

not completely overcome the inversion. (See Figure 13.13.) Under such conditions the pollutants mix more uniformly through the layer below the inversion. Unstable air and strong winds are inimical to formation of dense smogs. Rising air currents carry wastes upward and winds disperse them through a large volume of air. Prevailing winds carry pollutants away from single sources and produce leeward plumes that experience a progressive decrease in intensity of pollution. (See Figure 13.14.) Local winds, such as sea or valley breezes, may have a ventilating effect, but they may also act perversely, even returning polluted air that has drifted away from its source.

The effect of precipitation of local atmospheric pollution is not as great as might be expected. Although some materials are washed out by rain, wind and unstable air are more effective in cleansing the air during a storm. On a larger scale, however, much of the material that is dispersed into the air must return to the surface in precipitation.

The atmosphere is an amazingly effective sewage disposal plant, but it can be overtaxed locally. Factory location can often take advantage of variations in the ability of the atmosphere to disperse pollutants. Areas having a high frequency of low-level inversions are particularly unsuitable for factories that eject large quantities of airborne wastes. High chimneys help to disperse effluents above the surface level, but their effective height is limited by both construction difficulties and atmospheric conditions. The contaminants of some industries could be more efficiently dispersed by piping them to a hilltop during periods when stable air prevails. Pollution can also be alleviated by restricting the release of wastes to periods of favorable meteorological conditions, that is, strong winds and unstable air. Forecasts of *air pollution potential* greatly aid in controlling emissions during critical periods. (See Figure 13.15.) Mean depth of the surface mixing layer and mean wind speed within the layer are major criteria for forecasts. Areas of greatest pollution potential coincide with regions of frequent stagnating anticyclones and low-level inversions.

Figure 13.14
Smoke released from a meteorological tower indicates winds moving from opposite directions. Towers such as this aid studies of atmospheric pollution and other industrial problems. (Brookhaven National Laboratory)



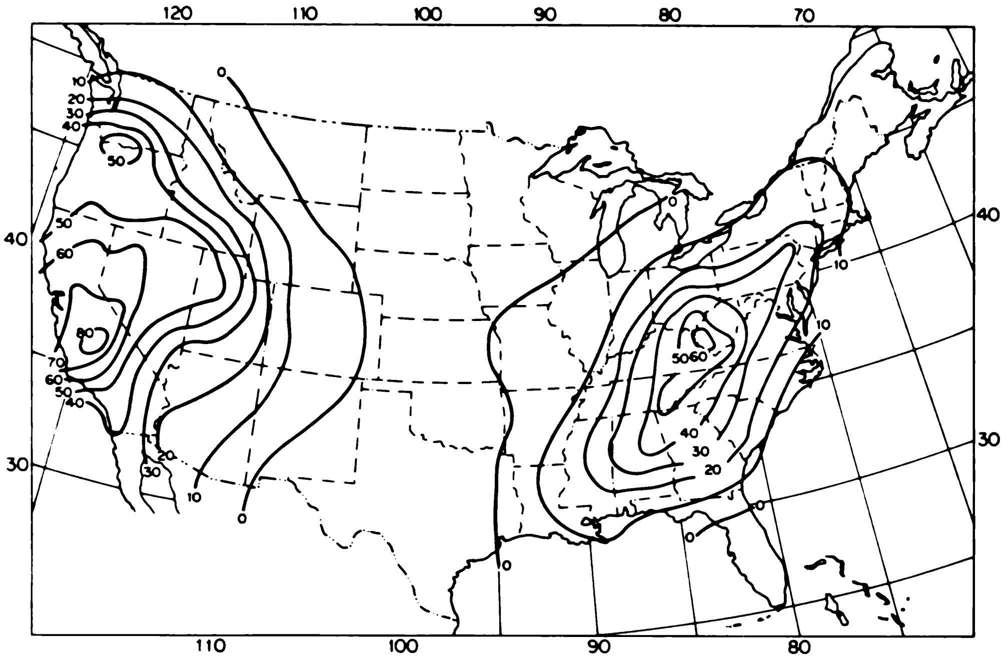


Figure 13.15
Predicted high pollution potential days per year in the contiguous United States. (After McCormick, Leighton, and M. Smith)

CONSTRUCTION

There are two broad aspects to the relationships between outdoor construction projects and climate or weather: (1) influences on structural design, and (2) direct effects on construction activities. A given structure must be designed to withstand stresses caused by climatic factors such as temperature fluctuations, wind force, humidity changes, or loads of snow and ice. Selection of materials is also influenced by climatic conditions; certain materials are better suited to climatic extremes by virtue of their inherent strength or their resistance to damage by chemical or physical weathering processes.

Temperature is one of the most important climatic elements to be considered in construction design. Metals, in particular, expand and contract with changing temperature, and allowance must be made to prevent buckling of girders and supports during warm periods or contraction and possible breaks in cold weather. Steel cables sag in warm weather and become taut in cold weather. Vertical metal towers expand more on the sunny side than on the shaded side, thus undergoing a diurnal bending that creates strains. Paints which reflect a large part of insolation help to lessen this problem. Reflecting paints and safety devices are provided for storage tanks to minimize the danger of overflow or bursting due to expansion of the contents. In general, extreme temperatures and sudden changes of temperature bring on the greatest troubles. Concrete, stone, and other types of masonry are subject to damage when temperatures drop rapidly to below freezing, especially if they are wet. The expansion of ice in cracks or in damp earth adjacent to masonry walls commonly causes structures to break up if they are not properly drained and reinforced.

Wind exerts a direct force on any structure in its path, and the maximum expected wind speeds are among the stress factors considered in designing bridges, towers, and buildings. Calculations based on climatic records and wind-tunnel experiments assist structural engineering. Bridges are often located across gorges or in river valleys where they are exposed to unusually strong, gusty winds. Suspension bridges are particularly vulnerable to wind, for they tend to sway and to develop destructive undulations. A notable example was the first Tacoma Narrows bridge at Tacoma, Washington, which was destroyed by wind on November 7, 1940. On December 28, 1879, sections of the Tay Bridge at Dundee, Scotland, were blown down by westerly gales. A train that was crossing the bridge at the time fell into the water, killing all 75 persons aboard.

High towers, even if supported by cables, sometimes develop a "whipping action" and collapse under the force of strong winds. The total force of wind on an object in its path results from the impact of air on the exposed side and the lower pressure on the leeward. It depends on the size and shape of the object and is proportional to the square of the wind velocity. The full force of wind is never felt, even by flat surfaces, because the air tends to flow around them. Streamlining of surfaces can appreciably reduce the effect of wind pressure. Venting in broad structural members allows the air to pass through with less force also. Solid structures incorporate considerably more strength in order to withstand wind. Additional stress is placed on hollow structures as a result of the difference in pressure between the air inside and the wind outside. Vents help to equalize the pressure.

Relative humidity is significant in its effects on construction materials, preservatives, and paints. Metals are generally more susceptible to corrosion under conditions of high relative humidity. Chemical and physical deterioration of paints is speeded up by moist air, and masonry construction also disintegrates more rapidly under moist conditions. Rainfall not only enhances the destructive influences of moisture but also produces direct physical effects on exposed surfaces. Heavy rains erode earthen structures and cause floods which undermine foundations or carry away ground structures. Dams, canals, piers, and bridges must withstand maximum flood stages. Accumulations of snow create extra weight on horizontal surfaces: additional supports may be required to prevent collapse. Potential accretions of ice on towers, bridges, and cables also call for safety margins in design. When heavily loaded with ice, such structures are more vulnerable to the wind.

In the engineering design of construction projects, climatic records are of great value in determining the extremes to which the finished structure may be subjected. Actual construction work is related to both day-to-day weather and the climate. Much construction is seasonal, especially in climates where there are marked seasonal weather fluctuations. Economic efficiency dictates that work be planned carefully to take advantage of favorable weather and to avoid loss of time or damage to materials and equipment in bad weather. Climatic records of precipitation and temperature thus aid in scheduling various types of work. Concrete and mortar are damaged if frozen before they set thoroughly. If the amount of newly poured concrete is small, it can be protected by coverings or artificial heating. Frozen earth impedes some types of construction. On the other hand, in the tundra or muskeg, heavy loads can be moved over the frozen ground with less difficulty. Most machinery does not function well in temperatures below freezing, and special problems arise from the necessity for antifreeze, heaters, and closed cabs on mobile equipment. Metal parts break more readily in extremely low temperatures. Temperature also affects the efficiency of workers on a construction project. Very high temperatures appreciably slow the pace of manual work among the unacclimatized and may lead to illness among men doing heavy physical work; low temperatures make more clothing necessary and thereby complicate manual tasks. Accidents are more numerous during periods of freezing temperatures or excessive heat than under normal temperatures. Housing and feeding of construction crews are also influenced by temperature.

Precipitation generally has an adverse effect on construction. Heavy rains pit newly laid concrete, erode earth embankments, and slow down or stop most earth-moving activities. A deep snow cover usually brings work to a halt. Heavy precipitation and fog reduce visibility and thereby increase the danger of accidents. Special facilities have to be provided for protection of materials and equipment subject to damage by moisture. Winds tend to increase the unfavorable effects of precipitation as well as the accident hazard on outdoor construction work, especially on tall structures such as towers, bridges, or high buildings. Partially completed structures are generally more easily damaged by strong or gusty winds. Certain kinds of construction activities are specifically affected by wind. Spray-painting is an obvious example. Drying winds prevent concrete and mortar from setting properly unless their surfaces are specially dampened.

Additional Readings

Ackermann, William C., "Application of Severe Rainstorm Data in Engineering Design," *Bull. Am. Met. Soc.,* 45, 4 (1964), 204–6.

Battan, Louis J., *The Unclean Sky.* Garden City, N.Y.: Doubleday & Company, Inc., 1966.

Beckwith, W. Boynton, "The Effect of Weather on Operations and Economics of Air Transportation Today," *Bull. Am. Met. Soc.,* 52, 9 (1971), 863–68.

Buck, Robert N., *Weather Flying.* New York: The Macmillan Company, 1970.

Corey, C. P., "The Effects of Weather upon Electric Power Systems," *Bull. Am. Met. Soc.,* 30, 7 (1949), 239–41.

Finger, Frederick G., and Raymond M. McInturff, "Meteorology and the Supersonic Transport," *Science,* 167, 3914 (1970), 16–25.

Haltiner, G. J., and H. D. Hamilton, "Minimal-Time Ship Routing," *Journ. Appl. Met.,* 1, 1 (1962), 1–7.

Maunder, W. J., *The Value of the Weather.* London: Methuen and Co., Ltd., 1970.

Myers, Joel N., and John J. Cahir, "The Weather Business," *Weatherwise,* 24, 2 (1971), 64–67, 95.

Pack, Donald H., "Meteorology and Air Pollution," *Science,* 146, 3648 (1964), 1119–28.

Weather a Factor in Plant Location. Washington: U.S. Department of Commerce, Weather Bureau, 1961.

Webb, Willis L., "Missile Range Meteorology," *Weatherwise,* 16, 3 (1963), 100–107.

chapter fourteen

Climate and Housing

Human dwellings have reflected influences of weather and climate since prehistoric times for the excellent reason that the primary function of housing is to shelter the inhabitants and the material contents against the weather. Whether one has in mind a small cottage, a mansion, or a skyscraper, atmospheric conditions are relevant factors in efficient siting, choice of materials, design, and air conditioning of the structure. The task of "building climatology" is to analyze outdoor climatic factors that influence indoor climates. Temperature, radiation, precipitation, and wind are the chief elements requiring attention. Humidity, cloudiness, and visibility may be criteria for certain decisions. Economic and social factors normally have a major influence on housing, but the principles of applied climatology afford a basis for many refinements in location and construction of buildings. Indeed, that which is economically feasible and socially acceptable in housing is determined in part by climatic circumstances.

CLIMATIC ASPECTS OF SITE

Assuming it has been decided to construct a building in a given city or rural area, local microclimatic conditions weigh heavily in the selection of the best site. A good building site may possess microclimatic advantages which offset some of the disadvantages of the prevailing regional climate, that is, the macroclimate. Variable microclimatic conditions are commonly induced by local relief, but they may also be brought about by landscaping, adjacent buildings, water bodies, and industrial wastes. Since some of these controls of microclimate are likely to be altered by expanding settlement, especially in a city, it is advisable to anticipate possible major changes in an environment which might drastically affect the microclimate. Ideally, a complete site study should include the microclimate along with such matters as bedrock, drainage, land cost, or proximity to services. Unfortunately, detailed instrumental observation of the microclimate requires time and is not economically feasible except for the most expensive structures. The alternative is to apply known principles of microclimatology and to make reasoned inferences with respect to the climatic conditions at a particular site.

Wind is the most important climatic element in site selection, for it produces direct effects on a building and modifies temperature and moisture effects. How much shelter from the wind is desirable depends on the regional climate and on the type of building on the site. In warm and humid climates, free circulation of air helps to lessen excessive humidity and high temperatures; in cold climates, wind may have value in moisture control, but it also increases convective cooling. In either case, high-velocity winds are to be avoided if practicable, otherwise they must be allowed for in building design. Local relief is the most significant as well at the most permanent of the common controls of wind conditions at a specific site. Windward slopes, summits, and plains are likely to receive the full force of surface winds. Sites where topographic barriers produce a constriction in air flow often have strong and gusty winds. Valleys and slopes are subjected to the effects of air drainage and, possibly, mountain and valley breezes. These local movements of air can be of immense value in warm climates and in summer, but they may be annoying in winter. Selection of a site should take into account the directions of prevailing winds and their temperature and moisture characteristics, but a wind which occurs only a small percentage of the time may be more objectionable than all others combined. An outstanding example is a wind from the direction of sources of air pollution. The problem is not confined to cities. Few farmers would want their homes to the leeward of livestock pens or barns. Objectionable odors, pollens, dust, and even insects are carried by the wind.

Types of surface and ground cover to the windward modify wind force, temperature, and humidity. A forest, orchard, or park has a moderating effect on summer temperatures, but pavements and other bare surfaces magnify temperature extremes. Buildings in the vicinity modify both direction and speed of the wind and may create undesirable local currents. A cluster of large buildings absorbs heat during a warm day and radiates it at night. Summer nights may be abnormally hot in the area of its influence. Water bodies aid in reducing diurnal and seasonal temperature ranges and generate breezes. If they are too shallow, however, they warm readily and thus create oppressive relative humidities in their vicinities in summer. In winter in the middle and high latitudes, shallow lakes and ponds may freeze, rapidly losing their moderating effect.

Temperature conditions, other than those directly influenced by the exchange of air between the site and its surroundings, are controlled by sunshine and by elevation. Since the approximate location of the housing site is assumed to have been decided, latitude and its effects on duration of possible sunshine and noon angle of the sun are not variable factors. Average cloudiness may differ within a given area because of relief features, and industrial smoke may greatly decrease the effective insolation, another reason for selecting a site to the windward of industrial establishments. Maximum benefit from insolation is achieved by choosing a site on the slopes facing toward the equator, where the sun's rays are more nearly perpendicular to the ground surface. In contrast, poleward slopes receive the sun at low angles and may actually be in the shade for part of the day. In the Northern Hemisphere, slopes on the east and southeast are sunniest in the morning; west and southwest slopes are sunniest in the afternoon. Where maximum heating is desired, westerly exposures are better than those on the east because insolation is received at the time of highest air temperatures. An easterly exposure provides heating in the early morning; direct insolation is less in the afternoon, and cooling proceeds faster in the evening. Sites in deep valleys have a shortened period of possible sunshine, which may be a distinct disadvantage in winter. In a built-up area the difference in sunniness on opposite sides of a street is worth considering. Nearby tall buildings may shut out most of the sunlight.

The effect of elevation on temperature-design relationships is closely related to relief. Except where differences of several tens or hundreds of meters in elevation are involved, higher sites are cooled more by the freer circulation of air than by the effects of the normal lapse rate of temperature in the daytime. Air drainage and temperature inversions commonly disrupt the normal lapse rate at night. Under clear skies, a broad valley has greater extremes of temperature than adjacent slopes, where a temperature inversion provides warmer night air. Slope sites at or above the average level of inversions have the advantage of being above most of the associated evils such as smoke, haze, and radiation fog.

Precipitation differences will not be greatly significant in the normal range of site possibilities, unless there are wide variations in elevation and orographic conditions, in which case lee slopes will usually be drier. If there is considerable wind, some knolls and ridges produce the opposite. It is never wise to build on or near a river bank without thorough deposited on the lee. An examination of the vegetation often aids in determining whether this is a common occurrence. Sites exposed to the wind will be most affected by driving rains. The pattern of snowdrifts depends on local eddies produced by obstructions. If deep snow is a winter problem, a site near a highway or street which is regularly cleared is better than an isolated one, unless one prefers to be snowbound. Where heavy showers occur or where there is a heavy runoff of snowmelt, flooding can necessitate excessive costs in design and maintenance of a building and its site. It is never wise to build on or near a river bank without thorough knowledge of the precipitation characteristics of the watershed and the resulting flood regime.

CLIMATE CONDITIONING

Even the best of building sites will be subject to weather changes, and other factors may make it necessary to use a site which is poor from the point of view of applied climatology. Nevertheless, proper orientation, design, and choice of materials for a house can overcome many of the climatic disadvantages of a site, the primary objective being to create a suitable climate in and around the house, that is, *climate conditioning.* Climate conditioning is concerned with landscaping and the placement of other buildings as well as with the design of the house itself. In its broadest sense, it includes certain aspects of the field of city planning, for groups of buildings and the associated streets and parks create their own microclimate. For many householders, convenience, architectural harmony, view, and cost of maintenance offset the minor advantages of a slightly better microclimate, but it is well at least to recognize the climatic factor in choosing a housing site.

BUILDING ORIENTATION

The effects of weather elements are strongly directional. Orientation of a building with respect to wind and sun consequently influence wind force, precipitation, temperature, and light on different external surfaces. Narrow building lots in a residential area preclude wide choice in orientation of a dwelling. Adverse climate requires compensation in design and in the mutual protection afforded by closely spaced houses. Where there is greater freedom, it is expedient to orient buildings to achieve the maximum control over the microclimate. The broad face of a house obviously endures greater wind pressure than does the narrow dimension under the same wind speeds. This may be desirable if high temperatures are prevalent, but it becomes a problem if winds are exceptionally vigorous or cold. Strong, cold winds on the side of the main entrance are particularly objectionable. Orientation with respect to both local and prevailing winds as well as to buildings or other windbreaks in the vicinity determines the relative effect of wind on different parts of a house. A wind rose showing average windspeeds and frequencies from different directions is especially useful if it is interpreted in connection with temperature and moisture data. The possibility of wind-driven rain merits special investigation.

The rules which govern the relation of sunshine to site exposure apply equally to building orientation. In the Northern Hemisphere, the greatest amount of insolation is received by the east and southeast sides of a house in the morning and by the west and southwest sides in the afternoon. In summer, with increasing latitude, the path of the sun in the sky describes more complete circles so that the sun rises north of east and sets north of west; in winter the arc of the sun's visible path is shortened, and east and west exposures receive very little sunlight.

CLIMATE CONDITIONING THROUGH DESIGN

The climatic factors related to housing design are insolation, temperature, wind, and moisture. Insolation provides heat and light and has certain health-giving powers. Climate conditioning seeks to control its effects through such design features as shape, layout of rooms, placement of openings, insulating materials, overhangs, or roof orientation and slope. The functions of various rooms determine the most efficient layout with respect to the sun. For example, a kitchen and breakfast nook are best placed where they will receive morning sun, whereas the living room may benefit from afternoon sun, especially in winter. Bright sunlight can sometimes be a nuisance; the study and certain kinds of workrooms are better located on the north side of the house, where lighting is more even throughout the day.

By varying the ground plan from the traditional rectangle, it is possible to obtain considerably more sun control for various rooms. An L-shaped house can be oriented to provide a greater variety of sunny exposures than a square one covering the same ground area. Admission of sunlight to the interior need not be confined to windows in vertical walls on the sunny exposures. Skylights offer one method of increasing the light in a one-story house or on the top floor of a multistoried building. They are

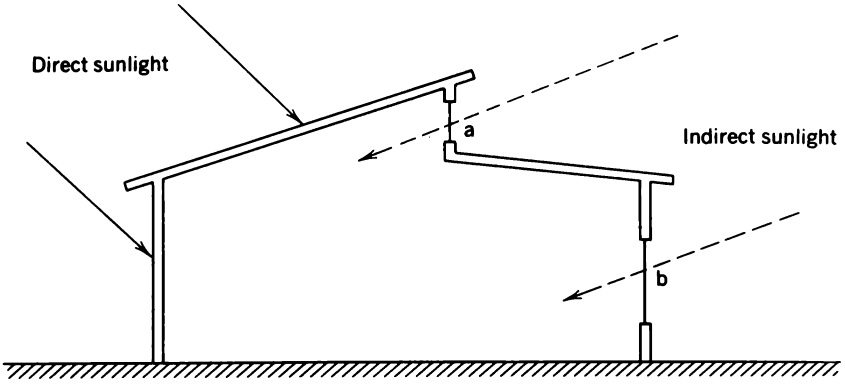


Figure 14.1
Principle of the clerestory. The clerestory window at "a" and the wall window at "b" admit scattered and reflected light. The opaque roof and wall on the sunny exposure keep out direct sunlight.

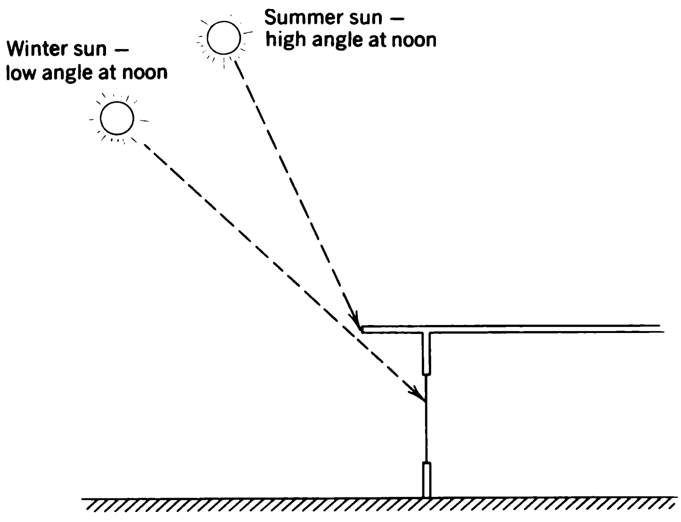


Figure 14.2
Effects of an overhang in winter and summer.

difficult to maintain, however. Clerestories are suited to rooms on the shady side of a building where sunlight is desired. The principle of the clerestory is illustrated in Figure 14.1. In combination with ceilings of the proper color and texture it can greatly improve lighting of northern rooms.

Unwanted bright sunlight is controlled from the interior by means of curtains or blinds over openings. Glass bricks and translucent windows also reduce glare while admitting light. Overhangs, awnings, and screens serve much the same purpose on the outside and have the added advantage of reducing heating of walls. In sunny climates of the middle latitudes overhangs can be designed to expose walls and windows in winter when the sun is at a low angle but to provide noonday shade in summer. (See Figure 14.2.) The exact size of overhangs for the degree of shading desired is calculated from the noon elevation of the sun at different times of the year for the latitude of the site. In low latitudes, overhangs shade walkways and the ground around the house, thereby helping to reduce the temperature of the walls. Overhangs would have to be inordinately large in high latitudes to afford shade which is not usually necessary anyway.

Design features in a well-built house take into account possible extremes in outside temperature and aim to maintain indoor temperatures at a comfortable level with a minimum of artificial heating or cooling. Basically, this entails control of the processes of heat transfer: radiation, conduction, and convection. These processes are affected by precipitation, humidity, and wind as well as by insolation and air temperature.

Solar heat is a blessing or a problem in relation to a specific house design depending on the climate and the season, and it is sometimes difficult to separate the benefits from the disadvantages. Well-insulated walls and roofs inhibit the passage of heat by conduction to or from the interior of a house. Light-colored and smooth surfaces reflect a greater percentage of sunlight than do dark and rough surfaces and consequently do not heat so rapidly under intense insolation. Through selection of materials for walls and roofs or the paint applied to them, it is possible to control solar heating to some extent. A light-colored surface will also reflect insolation in winter, but since there is less sunshine at that time this is generally not so important as summer cooling. Double roofs between which air can move freely are effective in reducing the flow of absorbed solar heat into a house in hot climates. Venting of attics has a similar though less marked effect. In semiarid and arid climates, flat roofs can be kept covered with water and thus aid cooling by evaporation. Short-wave solar radiation passes through clear air and windows with relative ease (even at low air temperatures) and upon being absorbed is converted into long-wave radiation which is partially trapped. Single-panel windows conduct and radiate a great deal more heat than insulated walls, however, and the gain in solar heat through large and numerous windows during short, sunny winter days may be more than balanced by losses at night or on cloudy days. Storm sash or double-pane glazing with the intervening space sealed airtight will reduce conductive loss. Cooling due to evaporation of water from the exterior window surfaces can also be reduced in this way. Drapes drawn over the interior of windows partially curb radiation losses. In general, the less sunshine at a particular site, the more glass is needed to overcome dark interiors, but it must be remembered that glass expedites the passage of heat in both directions.

Wind causes convective cooling, intensifies precipitation effects, and exerts direct force upon a house. Therefore house design should provide for insulation against too much cooling, tight construction to prevent the entrance of cold air or wind-driven precipitation, and overall strength to withstand wind pressure. (See Figure 14.3.) Under hot conditions, a movement of air through a building is useful in controlling temperature and humidity. Tropical houses have numerous openings to permit free flow of air. Native dwellings are commonly designed so that walls can be propped up or rolled up all around in a manner to provide shade but allow breezes to circulate. The tent of the desert nomad is a similar example of climatic conditioning through design. The flaps can easily be arranged to create shade and admit air by day and to restrict cool breezes by night (see Figure 14.4). Movement of air through a house

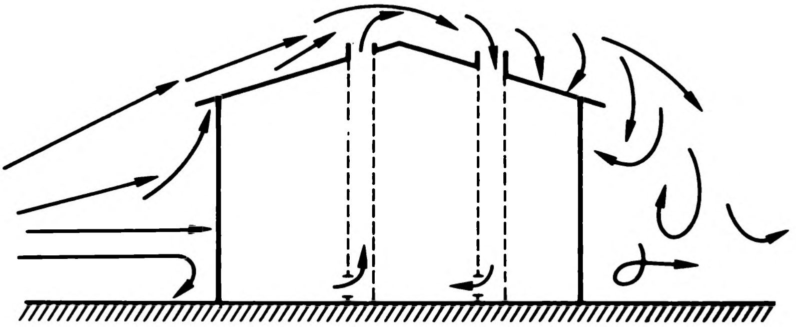
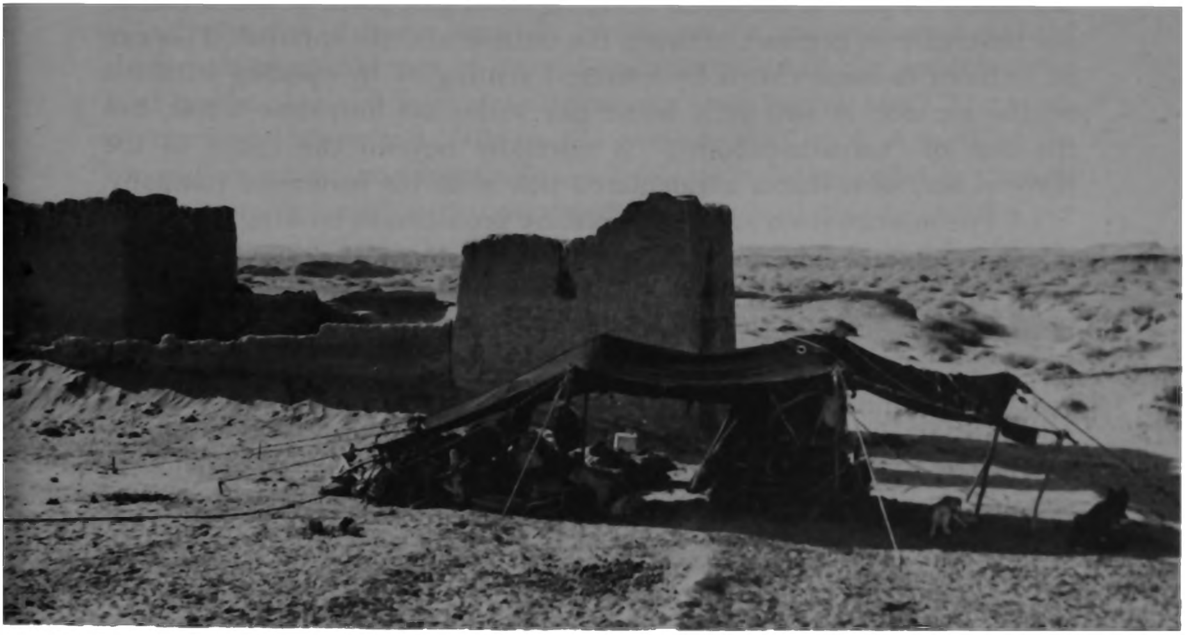


Figure 14.3
Typical wind effects on a house. Eddies generated across the roof create unequal pressures and gustiness; they may cause reverse flow in chimneys or ventilation ducts. Open windows further modify air flow both inside and in the vicinity of the building.

Figure 14.4
Climatic conditioning through design in a tropical desert. (An Exxon photo)



is modified by the openings and obstacles as well as by overhangs and the shape of the exterior. Various kinds of deflectors aid in directing air currents. For maximum airiness, floor-to-ceiling walls may need to be replaced by partial dividers or screens, for to have its maximum effect a breeze must pass through the house without being trapped in rooms on the windward. Some architects experiment with scale models of buildings in a wind tunnel before making final decisions on design details.

The design that allows free air movement may also admit precipitation carried along by the wind. Overhangs, louvers, and other types of deflectors help to overcome this. The attic venting that invites cooling air in summer provides an entry for rain or snow if not closed in the wet season or winter. Protection of the walls from precipitation is accomplished by overhangs except in the most violent storms. Walls need to be of tight construction and joints closely knit where there is danger of precipitation being blown into small openings. Even if it does not penetrate to the interior of the house, moisture will speed up deterioration of the outside walls. Canopies over entrances protect doors and shield the opening when it is in use. Vestibules extended beyond the main wall serve the same purpose and are more efficient in cold climates.

It was pointed out in the discussion of orientation of a house that the larger the surface exposed to the wind, the greater will be the total wind pressure upon it. Long eaves and overhangs tend to catch the wind and therefore increase the danger of wind damage. Loosely attached roof coverings of any type can be lifted and swept away by a fierce wind. Low, streamlined buildings of simple design are best for regions with violent winds. Departures from this principle call for compensating strength in structural members and walls. Possible destructive effects of flying debris to roofs, walls, and windows must also be taken into consideration. A part of the stress resulting from gale force winds is due to the difference in pressure between the outside and the interior. This can be reduced to some extent by louvered venting or by opening windows on the lee side. A well-built house can withstand hurricane winds, but the cost of "tornado-proofing" is normally beyond the range of the homeowner, who shares a calculated risk with his insurance company.

The most obvious aspect of moisture problems in housing is precipitation. Roofs and eave troughs should provide for rapid removal of heavy rains and for disposal of the runoff without damage to the surrounding grounds. If not properly diverted, runoff, carrying dirt, soot, or loose paint, may stain walls. Where there is danger of seepage along basement walls and foundations, tile drains will facilitate drainage away from the building. Water which enters crevices and subsequently freezes can be particularly destructive. Overhangs protect entrances, paths, and windows but must be designed with sun and wind in mind as well. Windows that

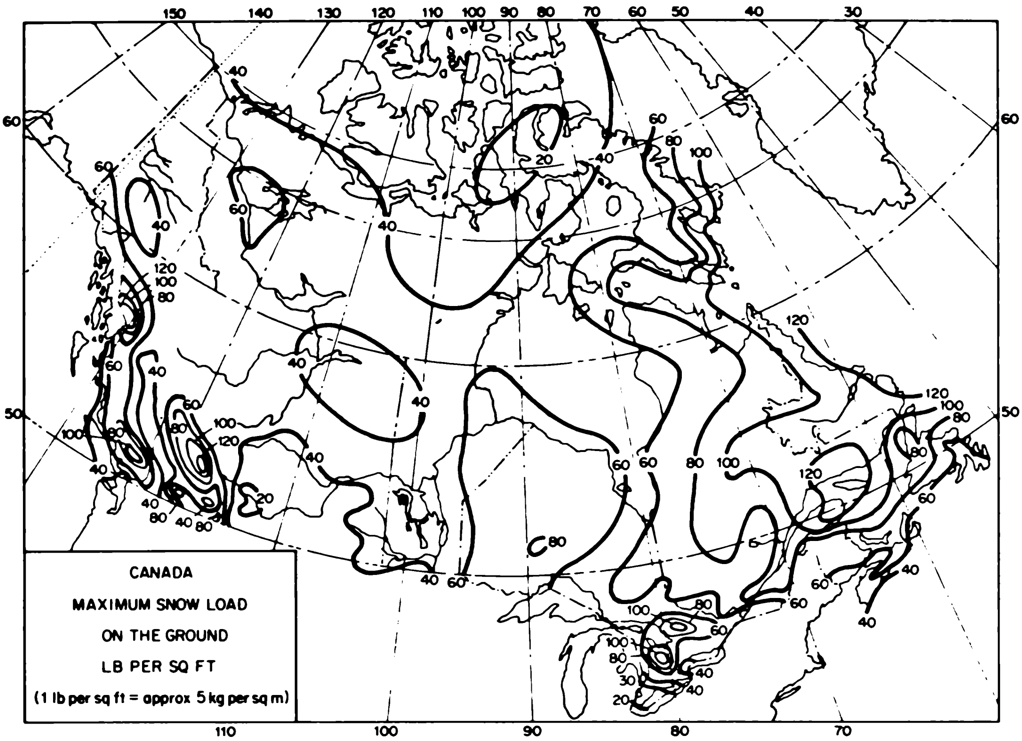


Figure 14.5
Generalized distribution of maximum snow load on the ground in Canada. Values incorporate maximum expected snow weight and the weight of maximum expected one-day fall of rain. (Data from Canadian National Research Council, National Building Code)

are placed too low are subject to rain splatter and covering by drifted snow; doorsills need to be high enough to hold back a shallow flow of water which is likely to accumulate on a doorstep in a deluge.

In regions of heavy snowfall, roofs must be strong enough to hold the weight of several tons of snow. Some indication of the possible snow load can be obtained from climatic records. (See Figure 14.5.) The stress is augmented when rain falls onto a snow-covered roof. A method for computing maximum snow loads as a basis for building codes in Canada uses maximum anticipated snow depths, an assumed density of 0.2 g per cm^3, and an added weight representing a one-day maximum fall of rain. A steeply pitched roof will permit wet snow to slide off, easing the stress on the roof but creating a hazard beneath the eaves. Gables and cornices tend to collect drifts on the leeward, producing an uneven weight distribution. Gentler slopes hold "dry" snow, which has valuable insulating properties in extreme cold.

To obtain benefit from a cover of snow, the roof itself must be well insulated so that the snow is not melted by heat from the house. Otherwise, ice will form, reducing the insulating capacity. Moreover, melt water which flows down the roof is likely to form icicles along the eaves; adequate drainage of melt water is thus impeded and the danger of falling icicles is introduced.

Hail damage is usually greatest to roofs and windows. Roofing materials should be resistant to the pounding of hailstones. Shutters, or preferably heavy screens, afford direct protection to windows, although overhangs usually can prevent hail from striking windows beneath. Skylights are inadvisable in areas subject to hailstorms.

The influence of fog on design is closely associated with sunshine. In addition, frequent fogs enhance the danger of moisture damage to building materials and paint. Where air pollution is common, its chemical effects are greater when associated with fog, dew, or frost. Streaking of walls often results from precipitation of dirty fog particles. The choice of paint should be made with these factors in mind.

Lightning occurs primarily with thunderstorms; where they are common lightning protection should be an integral part of house design. A stroke of lightning incorporates both an electrical discharge and intense heat that may start a fire. The principle of protection involves provision of a direct and easy passage of the current to earth. Metal lightning rods attached to the highest parts of the roof and heavy cables to conduct the current to the ground serve this purpose. Utility wires, antennas, and other wiring attached to the building should be grounded, preferably on the exterior. Building codes usually include regulations for lightning protection.

AIR CONDITIONING

Rarely can site selection, orientation, materials, and design create the desired indoor climate at all times. Air conditioning is the last resort in the overall attempt to provide a suitable indoor climate. In popular usage, the term has sometimes been restricted to the artificial cooling of the interiors of buildings. In the broader sense used here, it includes all attempts to modify indoor temperature, humidity, air movement, and composition of the air by artificial means. The demands placed upon air conditioning in the control of these elements depend on the outdoor climate, building design and its related factors, and the kind of indoor climate which is wanted. Though many of the principles remain the same, air conditioning of hospital operating rooms is quite a different problem from refrigerating a cold-storage locker. In residential buildings, the functions of different rooms influence requirements for comfort. Most important, individuals differ widely in their perception of comfort.

HEATING

Temperature is unquestionably the most important element of indoor climate. In the past much more attention has been given to heating than to cooling, for technology has flourished in the middle latitudes, where cold winters have been a greater problem than hot summers. Merely raising the air temperature in a cold room does not meet all the requirements for a comfortable environment; humidity and air movement are additional factors affecting the human response to air temperature. The concept of *effective temperature* combines the effects of air temperature, humidity, and air movement. Past studies of human comfort have defined effective temperature as the temperature of nearly calm, saturated air which would elicit the same thermal sensation for a normally clad, sedentary person as that produced by the actual dry-bulb temperature, relative humidity, and air movement. Indoor air rarely is saturated, however, and a relative humidity value that approximates average conditions affords a more realistic standard for comparison. The American Society of Heating, Refrigerating and Air-Conditioning Engineers has adopted a modified definition that incorporates 50 percent relative humidity rather than saturation as the basic humidity value. Figure 14.6 is the resulting ASHRAE Comfort Chart. It shows combinations of wet-bulb and dry-bulb temperatures, as well as relative humidity values, that elicit corresponding effective temperatures for persons who are adjusted to indoor conditions where the speed of air movement is less than 0.23 meters per second (45 feet per minute). An increase in air movement lowers the effective temperature; an increase in relative humidity raises it.

Because the effective temperature varies among different people, a composite is derived from the individual reactions of persons in a group. The shaded comfort zone in Figure 14.6 indicates the range of temperature and humidity conditions which most office workers find comfortable when wearing normal indoor clothing. The partially overlapping diamond-shaped area is the comfort zone for lightly clad persons at rest. The effective temperature lines and comfort zones are based on the responses of American subjects in controlled laboratory experiments. They do not necessarily apply in all climatic regions nor among groups of people who are accustomed to other indoor conditions. For example, in some mid-latitude countries where room temperatures are generally lower the optimum effective temperature for comparable levels of activity may be appreciably lower than in the United States. In the tropics comfortable effective temperatures are higher owing to different patterns of climatic experience and customs in dress and diet.

Heating systems that depend on forced circulation of air to distribute heat must be appraised in terms of effective temperature rather than the dry-bulb air temperature alone. Cool air introduced into a room suffers a decrease in relative humidity upon being heated. Fortunately, the range of comfortable relative humidity is rather wide—between 20 and 70 percent—so that there may be no resulting discomfort unless the outside air is abnormally cold and dry. When the indoor relative humidity is too low water can be evaporated into the air, preferably under control in conjunction with the heating or ventilating system.

Another factor affecting the feeling of comfort in a room is the transfer of heat by radiation between the body and walls, floor, ceiling, and solid objects such as furniture. Radiational loss to cold walls may create a sensation of discomfort even though air temperature is at the theoretical optimum. The body may experience simultaneous gains and losses of radiant heat on different sides. The average effect is expressed as the *mean radiant temperature,* an index combining air speed and *globe temperature.* The latter is measured by an ordinary thermometer inserted in a blackened, air-filled, copper sphere about 15 cm in diameter. When wall temperatures are low due to cold weather, it is necessary to keep the indoor air temperature a few degrees higher than the standard in order to compensate for the radiative loss. Conversely, warm surfaces radiate heat to the body independently of the temperature of the intervening air and can maintain comfort, even though air temperature is two or three degrees below the standard. This principle is employed in panel heating, also known as radiant heating, which warms people and objects in the room by radiation from large surfaces (for example, walls and floors) having comparatively low temperatures of 30° to 55°C. Panel heating has the advantage that convective air movement is held to a minimum and heating is much more uniform throughout the room than when smaller, high-temperature heat sources are used.

The gradient between temperatures indoors and those of the outside air determines, to a large extent, what amount of heating is required, for even the best-designed house will lose heat to the exterior by conduction. Evaporation, the escape of heated air, and convective cooling by wind are also factors which need consideration in planning a heating system, however. Wind promotes both convective and evaporational cooling. Direction as well as speed is significant, since a house may be more vulnerable to these types of cooling on one side than on another. A house loses a part of its heat by radiation to cold ground surfaces, other buildings, and to the atmosphere. In the latter instance, clear skies will enhance radiative cooling, especially at night and in winter. Cooling of a house by evaporation occurs primarily during and after precipitation and is increased by low relative humidity as well as wind. Escape

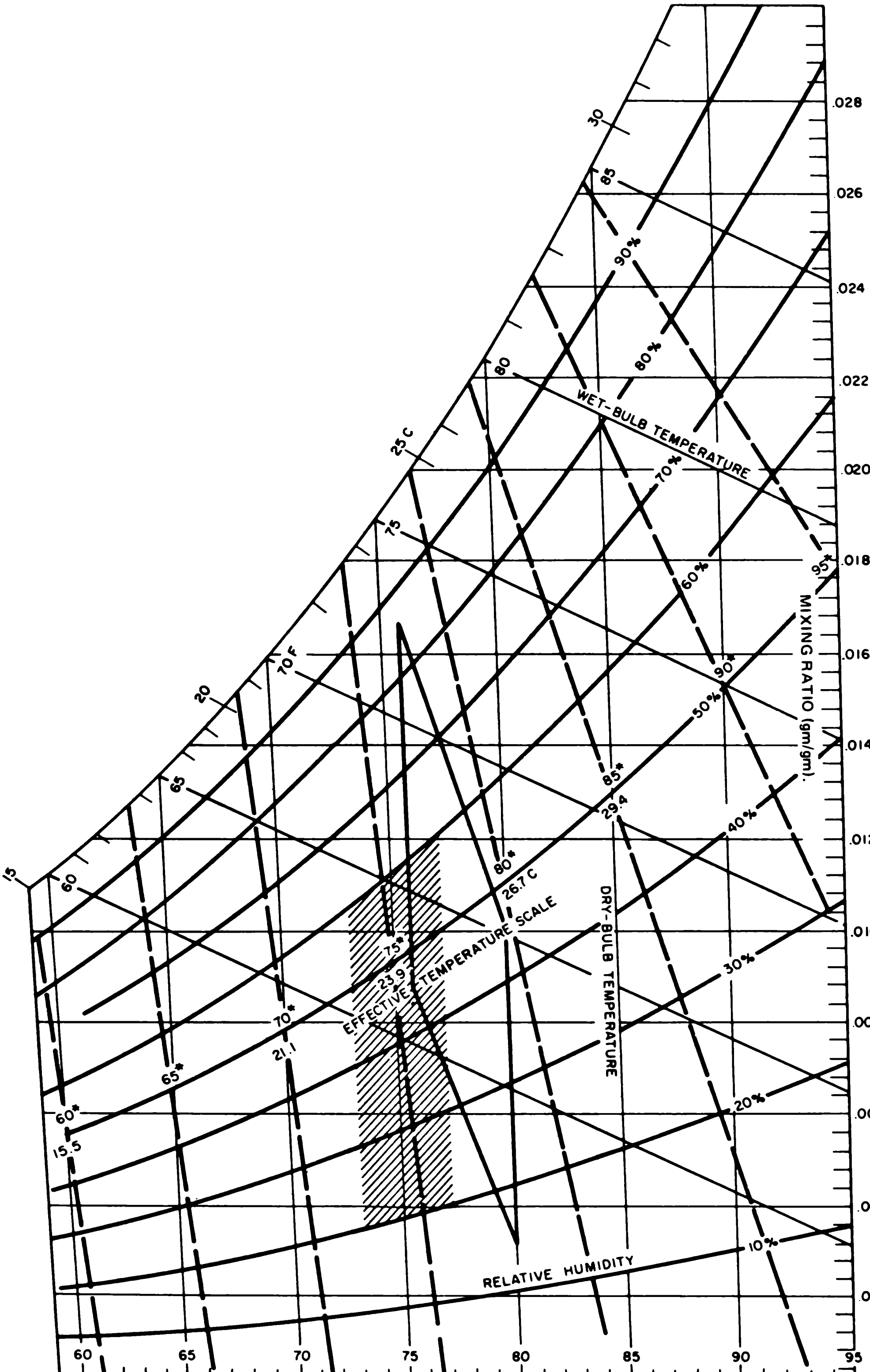


Figure 14.6

ASHRAE Comfort Chart. Dashed lines indicate effective temperatures based on 50 percent relative humidity and air movement less than 0.23 m/sec (45 ft/min). Shaded comfort zone is for normally clad office workers; diamond-shaped comfort zone is for lightly clad persons at rest indoors. (Reprinted by permission from *ASHRAE Handbook of Fundamentals 1972.*)

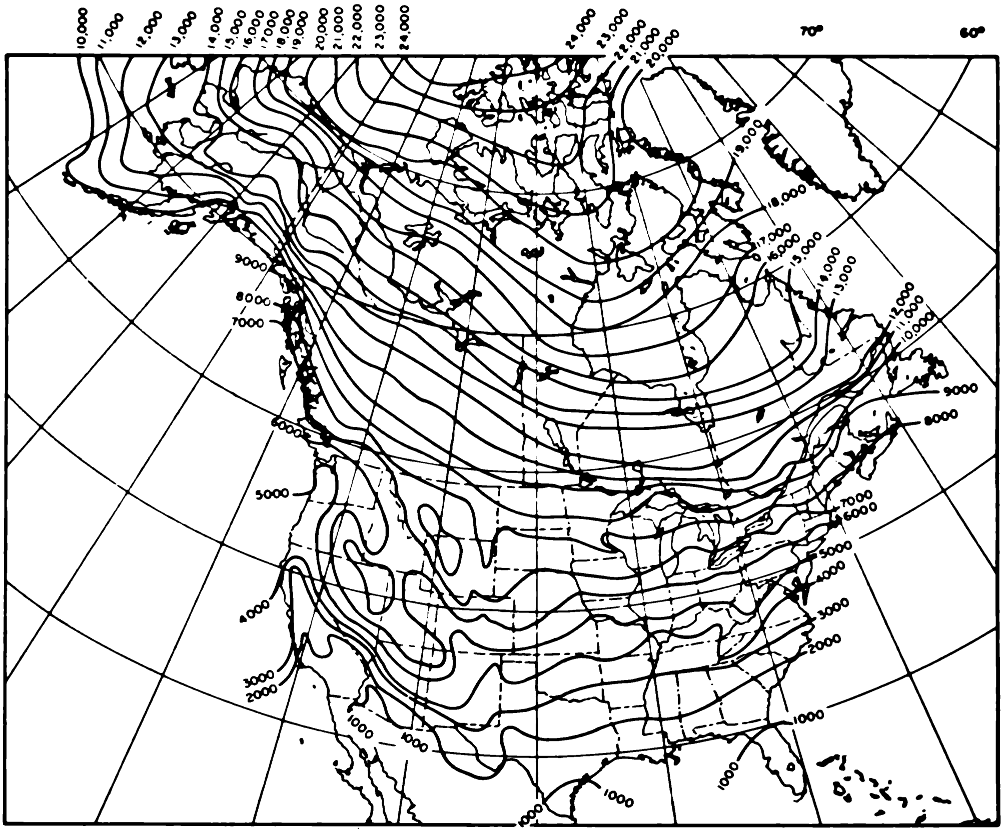


Figure 14.7
Average annual heating degree days in Canada and the United States. Base 65°F (18.3°C). (From United States Air Force, Geophysics Research Directorate, *Handbook of Geophysics,* published by The Macmillan Company.)

of heated air is controlled mainly by features of house design. Faulty ventilating systems often contribute to this loss. Fireplaces create a convection system which draws cool air into the room and carries warm air up the chimney.

Records of the pertinent climatic data are useful in determining the amount of heating required to overcome these cooling processes. Microclimatic observations in the immediate vicinity of a building are better than data from a weather station, which may be in a quite different situation. Nevertheless, regional temperature averages, ranges, frequencies of extremes, and heating degree-days are employed with some success in estimating the required output of a heating plant for a specific building. The term *heating degree-day,* widely used by heating engineers, is defined as a day on which the mean daily temperature is one degree below a certain standard. In the United States, the standard generally used is 65°F (18.3°C). Thus, a mean daily temperature of 60°F for a given day would yield 5 heating degree-days. Total heating degree-days for a month, season, or year are found simply by adding the accumulated degree-days for the period. Figure 14.7 is a map of the average annual heating degree-days over Canada and the United States. This index of heating requirements does not take into account cooling by radiation, wind, or evaporation. For more accurate computations, research is needed to develop an index which will combine all cooling factors.

Questions involving the type of heating system, kind of fuel or power, operating costs, and efficiency are primarily problems for the heating engineer, but they have climatic relationships. The design and capacity of a heating system should be such as to meet requirements imposed by weather changes and extremes. Ideally, automatic control mechanisms should have sensitive elements exposed to outdoor as well as indoor conditions in order to avoid a lag in adjustment of the indoor climate. The amount of fuel necessary to operate a system is estimated in terms of the heating requirements, which, in turn, are closely correlated with weather conditions.

COOLING

Under summer or tropical conditions, the main objective of air conditioning is to reduce the effective temperature. The practical methods include lowering air temperature, lowering relative humidity, and increasing air movement. Design and orientation of a building should be exploited to the fullest to accomplish these aims at the lowest expense. If the desired conditions cannot be obtained, as for example in interior rooms or under extreme temperatures, some form of cooling mechanism may be necessary. Fans speed up air movement and promote cooling by increasing evaporation of skin moisture. At night, cooler outside air can be drawn into a room by exhaust fans placed near outlets at high levels. Adaptations of refrigeration units placed at air intakes actually cool the air and are most effective if combined with fans and used in a room that is closed except for intake and exhaust openings. Comparatively efficient units of this type have been developed which are no more expensive than heating systems. If heat is combined with high relative humidity, the effective temperature may be lowered by passing the air through an apparatus which condenses a part of the moisture on cold coils or otherwise dehumidifies it. Air that is too dry can cause skin discomfort and irritate respiratory passages. In hot, dry climates air can be passed over water or through spray to increase its relative humidity and, at the same time, cool it by evaporation.

Cooling degree-days may be used as a basis for determining the amount of energy necessary to reduce the effective temperature of warm air. Although a cooling degree-day may be considered as a day on which the temperature is one degree above a desired base temperature, an expression incorporating a humidity value provides a more satisfactory index of the amount of cooling that will create comfortable conditions.

Under certain circumstances, an air-conditioning system may be called on to cleanse the air of pollutants. Dust, soot, pollens, and other solid materials are removed by means of filters, by "washing" the air in a spray chamber, or by precipitating the particles on electrically charged screens. Objectionable gases are much more difficult to remove, although some are soluble in water and can be partially controlled in a spray chamber. Chemical modification of gases is practical only under extreme conditions and should not be necessary in a planned residential settlement.

Additional Readings

ARONIN, JEFFREY E., *Climate and Architecture.* New York: Reinhold Publishing Corporation, 1953.

Building Climatology, Tech. Note No. 109. Geneva: WMO, 1970.

CONKLIN, GROFF, *The Weather Conditioned House.* New York: Reinhold Publishing Corporation, 1958.

GIVONI, B., *Man, Climate and Architecture.* Amsterdam: Elsevier Publishing Company Ltd., 1969.

Housing and Home Finance Agency, Division of Housing Research, *Applications of Climatic Data to House Design.* Washington: Government Printing Office, 1954.

LEE, DOUGLAS H. K., "Thoughts on Housing for the Humid Tropics," *Geog. Rev.,* 41, 1 (1951), 124–47.

OLGYAY, ALADAR, AND VICTOR OLGYAY, *Solar Control and Shading Devices.* Princeton, N.J.: Princeton University Press, 1957.

OLGYAY, VICTOR, *Design with Climate.* Princeton, N.J.: Princeton University Press, 1963.

PAGE, J. K., "The Effect of Town-planning and Architectural Design and Construction on the Microclimatic Environment of Man," in *Medical Biometeorology,* S. W. Tromp et al. Amsterdam: Elsevier Publishing Company, Ltd., 1963, pp. 655–70.

chapter fifteen

Climate and Human Comfort

Human health, energy, and comfort are affected more by climate than by any other element of the physical environment. Physiological functions of the human body respond to changes in the weather, and the incidence of certain diseases varies with climate and the seasons. Our selection of amounts and types of food and clothing also tends to reflect weather and climate. The state of the atmosphere even influences our mental and emotional outlook.

Different human beings do not react to identical climates in the same way, however; the relationship is complicated by individual physical differences, age, diet, past climatic experience, and cultural influences. Nor are all individuals equally adaptable to a change of climate. Of all life forms, nevertheless, humans are the most adaptable to varying atmospheric conditions. Climate has been blamed for human failure more often than the facts justify, especially in the tropics, but it is unquestionably a factor in human efficiency.

Among the climatic elements that affect the human body the more important are temperature, sunshine, and humidity. Wind exerts an influence largely through its effects on skin temperature and body moisture; the circulatory, respiratory, and nervous systems register changes in atmospheric pressure. Such elements as cloudiness, visibility, and storms induce psychological reactions that may be expressed ultimately as physiological disturbances. Acting together the climatic elements constitute the climatic environment that directly influences human comfort and well-being.

HEAT BUDGET OF THE BODY

The human body maintains a balance between incoming and outgoing heat by means of the chemical process of metabolism and the physiological processes of thermoregulation in response to external factors of radiation, temperature, moisture, and air movement. Each of the physical processes of heat transfer is represented in the exchange equation:

$$M \pm Cd \pm Cv \pm R - E = 0$$

in which M is the heat of metabolism, Cd the gain or loss by conduction, Cv the gain or loss by convection, R the gain or loss by radiation, and E the loss due to evaporation. If the metabolic heat exceeds the sum of the other budget elements, body temperature will rise; if the resultant is less than zero, body temperature will decrease. Normal rectal temperature is about 37°C (98.6°F); the temperature of the skin, which is the primary surface for heat exchange, varies around a mean of about 33°C. Metabolic heat depends ultimately on the intake and digestion of food. It is increased by muscular activity in the form of exercise or shivering. Excess heat is lost by radiation to surroundings, evaporation from the skin and respiratory passages, conduction to air and cold objects, convective transport in moving air (including cold air taken into the diaphragm and lungs), and minor losses to cold food and drink.

Within the normal range of air temperatures experienced by most people the most effective cooling is achieved by evaporation of moisture from the skin, and this is the type of cooling over which the body has maximum control. As activity is increased, so is the metabolic rate. The body adjusts by increasing the circulation of blood near the skin and by perspiration, although people differ greatly in their ability to sweat. As long as evaporation can remove the secreted moisture it will have a cooling effect, but if the relative humidity is high and the sweating profuse, a feeling of discomfort develops. Sweat which drops from the body merely represents a loss of water and is not effective in cooling the skin. The most comfortable values of relative humidity lie in the range of 30 to 70 percent. Increased speed of air movement along the skin surfaces aids evaporation. At temperatures below about 20°C, evaporation from the skin loses importance as a cooling factor, for the rate of perspiration is reduced. Radiation, conduction, and convection serve to reduce the temperature of exposed skin, but these processes are restricted by clothing. Heat loss from the lungs increases with rapid breathing (panting) and becomes critical as the air temperature drops to extreme minima. At −40°C the dissipation of body heat through the lungs may account for one-fifth of the total loss. As activity increases the volume of air inhaled also increases, and at very low temperatures damage to the lungs and a serious lowering of body temperature may result.

Air temperature as measured by a thermometer is not, in itself, a wholly reliable index of the temperature one feels. Individual responses depend not only on the elements of the heat exchange equation but also on subjective factors that vary widely among different persons and from time to time for the same person. The temperature the body actually feels, or senses, is known as the *sensible temperature.* It is not measurable by any instrument. Sensible temperature and wet-bulb temperature are approximately the same when the skin is moist and exposed to normal dry-bulb temperatures, however, since evaporation is then the principal cooling process. Thus, under warm conditions, low relative humidity tends to reduce the sensible temperature because the rate of evaporation is increased.

High relative humidity, in combination with high air temperature, results in a lowered rate of evaporation and consequently a high sensible temperature. The familiar saying, "It's not the heat; it's the humidity," is not strictly true, for the feeling of oppression which we associate with hot, humid weather is dependent fully as much upon the high temperatures as on the high relative humidity. When air temperatures are low, evaporation from the skin becomes secondary to conduction as a cooling process. Both air movement and high relative humidity increase the rate of conduction away from the body. The sensible temperature is increased in summer and decreased in winter by high relative humidity. But it is decreased in all seasons by increased wind speed, unless the air temperature is well above the body temperature.

Air temperature and a moisture factor have been the primary bases for an objective index of sensible temperature. Effective temperature is a useful approximation when air movement can be determined. (See the section on heating in Chapter 14.) Mean wet-bulb temperatures or mean humidity values show a close correlation with subjective body responses under a given set of temperature and wind conditions. They therefore afford a basis for determination of the range of conditions that are comfortable to most humans. Although the definition of comfort varies with individuals, living habits, physical activity, the seasons, and climates, nearly everyone finds conditions uncomfortable or oppressive when wet-bulb temperatures exceed 29°C (85°F). Dry-bulb temperatures below 16° to 18°C are experienced as cool or cold by everyone regardless of the relative humidity. E. C. Thom has suggested a further refinement of the temperature-moisture relationship at the warm margin of the comfort zone. In his formula

$$TH = 0.4(t_d + t_w) + 15$$

TH is the temperature-humidity index, t_d the dry-bulb temperature in degrees Fahrenheit, and t_w the simultaneous wet-bulb temperature. Comparable index values based on the Celsius scale result from the formula

$$TH = 0.4(t_d + t_w) + 4.8$$

For conditions of little air movement the index approximates the effective temperature (base at saturation). Most people feel discomfort as the index rises above 70 (F); everyone is uncomfortable when it reaches 79. The comfort zone for a group of people is a generalization derived from the individual comfort zones of the people in the group. Individual differences arise from variations in age, state of health, physical activity, type and amount of clothing, psychological factors, and past climatic experience. The normal, or group, reaction reflects not only the composite of individual responses but also the local climate and the season; hence the comfort zone (that is, the range of comfortable sensible temperatures for the group) depends on climate as well as the characteristics of the group. Because it is based on generalized statistics, it can tell little of the differing reactions of persons in different stages of acclimatization or of the effects of short-term weather changes.

COOLING POWER

Since the great variety of individual differences makes it impracticable, if not impossible, to obtain measurements of the sensible temperature the concept of *cooling power* has been developed to express the combined effects of air temperature and air movement in more objective terms. With respect to the human body, cooling power may be defined simply as the ability of the air to enhance the loss of body heat.

Several formulas have been devised to express cooling power in terms of observed values of temperature, humidity, and wind speed. The most-used indices of cooling power are derived from special instruments, however. Measurements of skin and body temperatures of living subjects under controlled conditions yield more realistic results and promise, eventually, to furnish the basis for sound conclusions on the reaction of the human body to varying conditions.

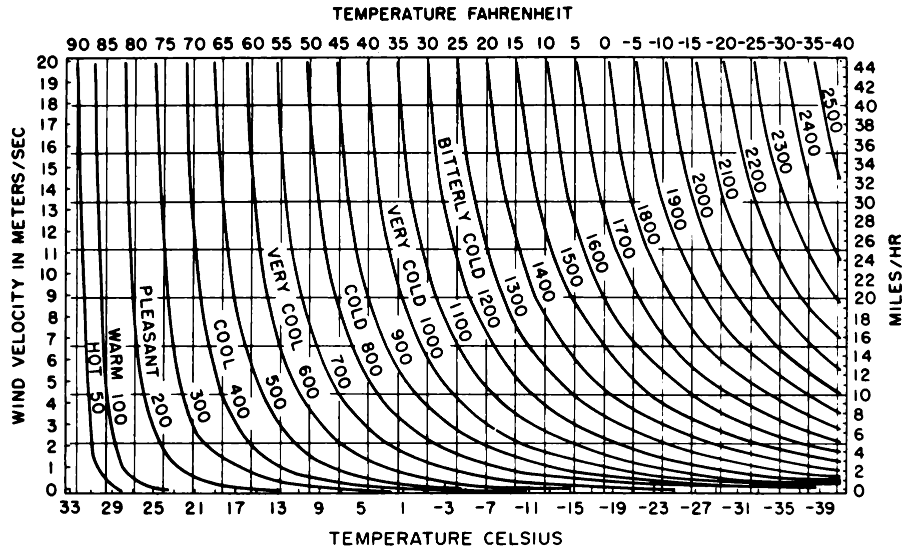


Figure 15.1

Wind chill chart. Curves show constant rates of cooling in kilogram calories per square meter per hour for different combinations of temperature and wind speed. To determine wind chill equivalents follow along cooling rate curves from one temperature-wind combination to another. Cooling rates are based on a body having neutral skin temperature of 33°C (91.4°F). When the dry convective cooling rate is less than the rate of body heat production excess heat is removed by evaporation of perspiration. Bright sunshine reduces cooling. The terms denoting relative comfort are based on subjective sensations of individuals at rest. (From United States Air Force, Geophysics Research Directorate, *Handbook of Geophysics,* Revised Edition, published by The Macmillan Company, 1960)

As we have seen, the body loses heat through a complex combination of several processes. Temperature and wind speed, which are the primary bases of theoretical cooling power, are only two of the factors in actual sensible temperature, but they are the major factors when the air is cold. Cooling effects due to low temperatures and wind have been called the dry convective cooling power of the atmosphere, or simply *wind chill.* (See Figures 15.1 and 15.2.) When air temperatures are above the minimum that the body can maintain by radiation and evaporation losses, moving air heats rather than cools the body. The colder the air and the higher the wind speeds, the greater is the loss by dry convective cooling. *Equivalent wind chill temperature* expresses the air temperature which in calm air would produce the same cooling effect as the actual temperature and rate of movement. For example, a 35-knot wind at 10°C elicits a chilling effect equivalent to a 1-knot wind at −15°C. The equivalent wind chill temperature of a 45-mile-per-hour wind at 20°F is about

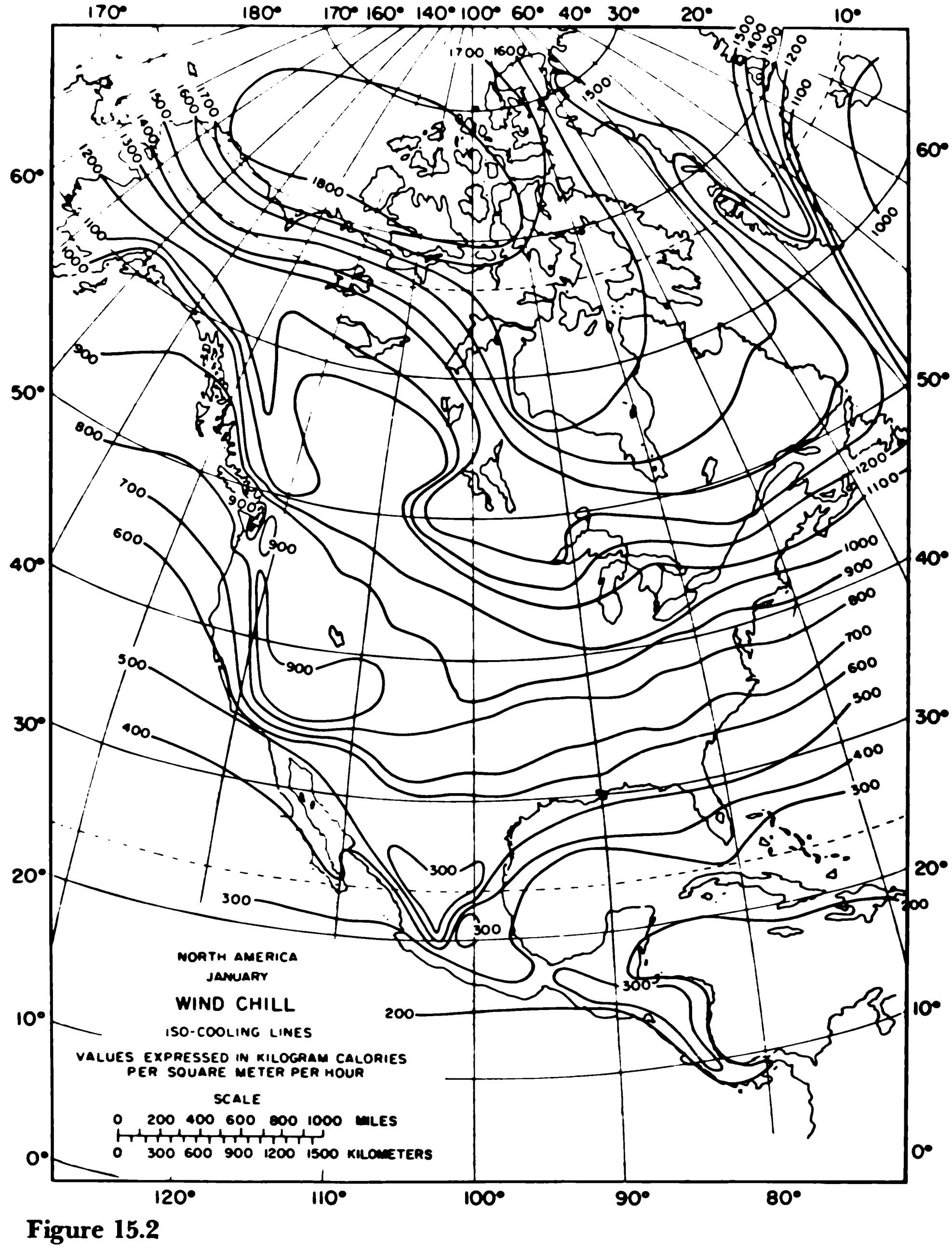


Figure 15.2
Wind chill values for North America in January. (From United States Air Force, Geophysics Research Directorate, *Handbook of Geophysics,* Revised Edition, published by The Macmillan Company, 1960)

the same as that of wind moving at 5 miles per hour with a temperature of −20°F. These relationships do not imply that the body will cool below the air temperature; rather they indicate a rate of cooling, which increases most rapidly with air movement at low wind speeds. (See Figure 15.1.) The sensible temperature nevertheless approximates the equivalent wind chill temperature. Were it not for the gales frequently encountered, the deterrent effect of low temperatures on high-altitude mountain climbing would not be nearly so great.

CLOTHING AND CLIMATE

With respect to climate, the main purposes served by clothing are: protection against temperature changes and extremes, protection from excessive sunshine, and protection from precipitation. Clothing is also worn to protect the body against physical damage such as abrasion, cutting, or burning. With respect to culture, clothing is worn for adornment, prestige, fashion, custom, and other social reasons which vary from group to group as well as among individuals. Very often, the design and selection of clothing is dictated more by cultural than by climatic factors.

Clothing protects against the cold by trapping still air within its open spaces and in the layer next to the skin. Layers of "dead air" are good insulators because they do not readily conduct heat away from the body. Therefore, the fundamental aim in design of effective clothing for cold weather is to provide insulation. Knit or loosely woven woolens are better than cotton cloth with a hard weave, and several thin layers of clothing are better than one heavy layer.

In order to be effective against wind chill, clothing must be reasonably impervious to the passage of air. Wind decreases the insulating capacity of clothing by forcing air into the openings and disturbing the still air. It also reduces the air layers between successive garments and between body and clothing by pressing the clothes against the body. Completely airtight materials are unsatisfactory because they do not provide for ventilation to remove moist air from near the skin. When clothing becomes wet from perspiration it loses much of its insulating power. In cold weather, wet socks are one of the first danger signs of possible freezing. This is one of the reasons for avoiding strenuous exercise in polar climates. Another is the fact that exercise sets up a bellows action in clothing, thus increasing convective heat loss. The ideal material for an outer garment in cold winds is one that will largely restrict the passage of wind but that will "breathe," that is, allow the escape of water vapor. Under-garments should be resilient enough to maintain their insulating properties against the pressure of the wind.

A rather obvious method of combating the effects of low temperatures and wind is the use of an electrically heated suit regulated by a thermostat. Heated suits have been used with some success by high-altitude fliers and certain sedentary workers, but they are generally impractical for persons active in the outdoors.

Special provision for protection of the eyes against snow blindness is necessary in cold climates and on mountain snowfields. Some form of cover is also advisable to prevent irritation of eyelids by the impact of blowing snow. Eskimos use goggles that have narrow slits to restrict the passage of both light and snow particles.

For protection against heat, clothing should be loose and allow free transfer of heat away from the body. One of the primary functions is to shade the skin from direct rays of the sun. Lightweight and light-colored materials that reflect insolation are best, and they need to be porous to allow maximum air movement adjacent to the skin. Wearing coats and ties in hot summer weather is justified only by fashion; in terms of applied climatology it is indefensible. For temperatures above that of the body, insulating materials are of some value in preventing overheating. The desert Bedouin finds the same clothing that shields him from extreme insolation in the daytime useful against the cold night air. But in highly humid air, such clothing would be unsuitable. In any case, water vapor in the air absorbs a large proportion of ultraviolet rays so that the danger from that component of the solar spectrum is not so great in humid climates. Head covers in hot weather are especially important to prevent illness due to excessive heat and sunlight. They must permit free ventilation of the head, shade the eyes, and reflect sunlight. Footwear to insulate against hot ground is also necessary in hot climates, especially if the ground is dry. Warm-weather shoes should be porous to allow as much air movement as possible.

Clothing to ward off precipitation is usually of special design and not necessarily effective against accompanying temperatures, which must be regulated by other clothing underneath. Completely waterproof garments made of rubber or oiled fabrics have the disadvantage of not allowing the escape of body moisture, and therefore become uncomfortable at any temperature. The so-called showerproof coats allow perspiration to escape but will not ward off heavy, continuous rain. A rain cape allows greater movement of air under its shelter but is more cumbersome when the wearer is working or encounters strong winds. Ideally, the material and the design of rainwear should prevent all rain or melted snow from passing through to the undergarments or body regardless of wind force and at the same time they should allow free escape of evaporated perspiration. For practical purposes, it is necessary to strike a compromise between the ideal and a utilitarian garment that will permit the maximum efficiency of physical activity. Rain clothes suitable for a policeman or sentry may be quite unsatisfactory for more active persons such as loggers. Although experiments in clothing design must take into account the variations and extremes of the weather elements, the type of activity performed by the wearer is also of major importance. There is no such thing as "all-weather" clothing. A great deal of discomfort is endured by persons who fail to modify the amount and type of their apparel to conform to changes in the weather or tendencies in their own heat budgets.

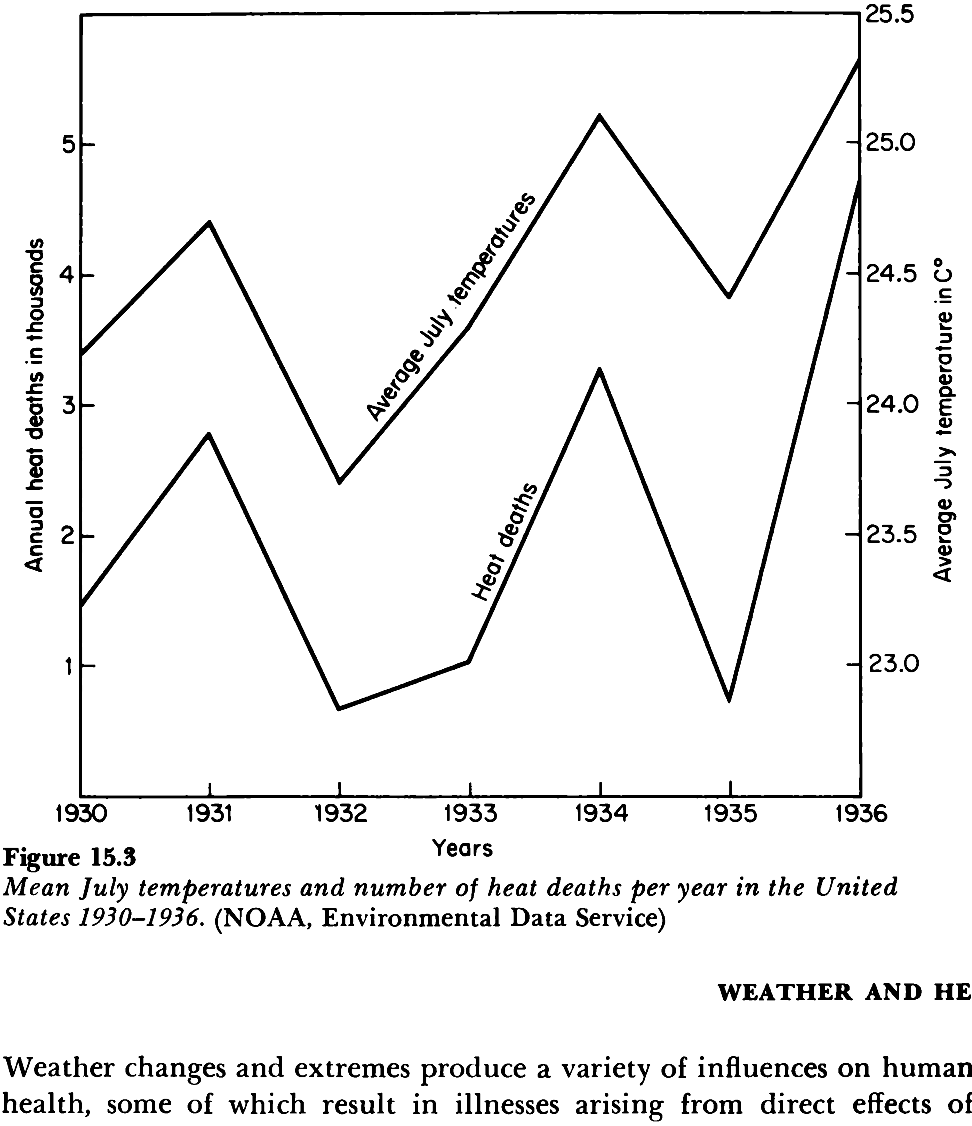


Figure 15.3
Mean July temperatures and number of heat deaths per year in the United States 1930–1936. (NOAA, Environmental Data Service)

WEATHER AND HEALTH

Weather changes and extremes produce a variety of influences on human health, some of which result in illnesses arising from direct effects of atmospheric conditions on the body. Temperature extremes are the most common causes of illness related to weather. *Heat stroke,* or *hyperthermia,* develops when the body is unable to maintain its heat balance at high relative humidity and air temperature above that of the body; it may lead to death if deep body temperature rises above the critical level of 42°C. Symptoms are fever, nausea, dizziness, and headache. Treatment entails reducing body temperature by means of cold baths. *Heat exhaustion* is a milder form of hyperthermia identified by dizziness, lassitude, and perhaps fainting. It is more common in crowded rooms than in the outdoors. When the body suffers an excessive loss of salts and water in perspiration, *heat cramps* may result. Adequate liquid and salt intake help to prevent heat cramps and also relieve the condition if it develops. Because of changes in metabolism and blood circulation, the appetite and digestion are impaired in hot weather; digestive disorders are more common in summer and in the tropics.

The summer heat waves of mid-latitudes create especially trying conditions in terms of health and comfort. The more disastrous heat waves accompany subsiding continental air; they affect large regions for several consecutive days and cause numerous heat deaths. Figure 15.3 shows graphically the correlation between average July temperatures and heat deaths in the United States during the drought years of the 1930s, when heat killed nearly 15,000 persons.

A common direct effect of low temperatures is *frostbite.* The extremities and exposed portions of the body are most likely to suffer, and the danger is heightened by increased movement. Perspiration, especially from the feet, accelerates conductive heat loss, as does the accumulation of condensed moisture about the face. The precautions against wind chill, namely, adequate clothing and moderation of physical activity, help to prevent frostbite. Prolonged exposure to low air temperatures can alter the body's heat balance, producing *hypothermia* and causing eventual death. The development of hypothermia is hastened by physical exhaustion. Low temperatures are an aggravating factor in a number of ailments such as arthritis, swollen sinuses, chillblains, and "stiff joints." Sudden lowering of temperature puts a severe strain on persons with cardiac disorders, although there is no evidence that climate directly causes such ailments. Indeed, virtually all physiological functions react in some manner to temperature changes.

Atmospheric pressure and relative humidity changes appear to bear a relation to certain kinds of pains, notably those associated with respiratory infection and muscular aches. The irritation of respiratory passages, which is commonly brought about by dry air, is further enhanced by wind and dust. Temperature and humidity are significant factors in the release of pollens and consequently affect the incidence of allergies. Very dry air is a major contributory cause of chapped skin and it inhibits the healing of sores and wounds.

To the extent that health is a function of nutrition, the influences of weather and climate upon diet indirectly affect human physiology. One facet of the influence results from the availability of different types of food in different climates. More important is the effect on appetite and selection of food. Temperature is the most significant consideration in this respect. Under cold conditions, the body requires a greater food intake to maintain heat; increased amounts of fats and carbohydrates are needed. Vitamins and minerals are, nonetheless, essential. Primitive peoples in the arctic achieve a balanced diet by eating virtually all parts of animals and fish. People of more sophisticated cultures secure a proper balance through a variety of foods. Nutritional diseases may result from insufficient intake of calories, vitamins, or minerals. Adequate water must also be provided, and where it is obtained from melted snow, the deficiency in minerals has to be made up through foods.

Nutritional requirements in hot climates differ from those in the mid-latitudes in that more salt, water, and certain vitamins are needed. Whether these requirements are met is another question. Suppression of appetite may be a major factor in malnutrition in the tropics. There is some evidence that meat and eggs produced in the tropics are deficient in certain nutrients, notably vitamin B_1, presumably as a result of the effect of heat on the living animals. Still other deficiencies in food can be traced to leaching of tropical soils by heavy rains. Whether humans can improve their heat tolerance by controlling their diet is as yet an unsolved problem.

SUNSHINE AND HEALTH

Besides the obvious relation between sunshine and air temperature, the solar spectrum produces several effects on the human body. Infrared rays are absorbed by the body or clothing and converted to heat, thus offsetting much of the cooling power of the air. It is therefore perfectly natural to seek the shade in hot climates and the sunny exposures in cool climates. The visible part of the spectrum (light) affects mainly the eyes. The intense sunlight of the arid tropics or that reflected off snowfields can cause forms of blindness, headaches, and related discomforts.

Ultraviolet rays are valuable for their ability to form vitamin D in the skin and to devitalize bacteria and germs. These qualities explain in part why many health resorts are in sunny locations. On the other hand, ultraviolet radiation can cause inflammation of the skin and sunburn to the point of illness. When the skin becomes pigmented (tanned) the pigment affords protection against further inflammation, but this adaptation differs greatly among individuals. Blonde persons are generally more susceptible to sunburn than those having darker skins. A "healthy suntan" is commonly regarded as the mark of a successful vacation, although it is probably no more significant than fresh air, a change of scene, and relaxation in promoting physical and mental well-being.

The apparent increase in the incidence of skin cancer with decreasing latitude may be caused by the greater amounts of ultraviolet radiation received where the sun angle is higher and the ozone content of the atmosphere is less. Combined with intense heat, ultraviolet rays are also a factor causing cataract of the eye.

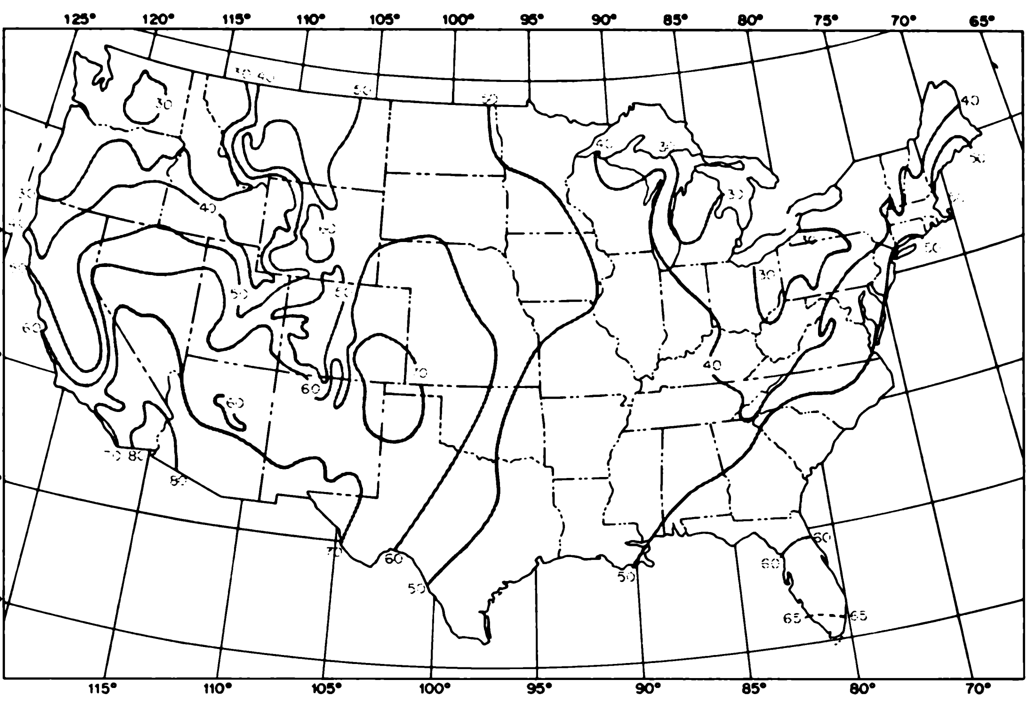
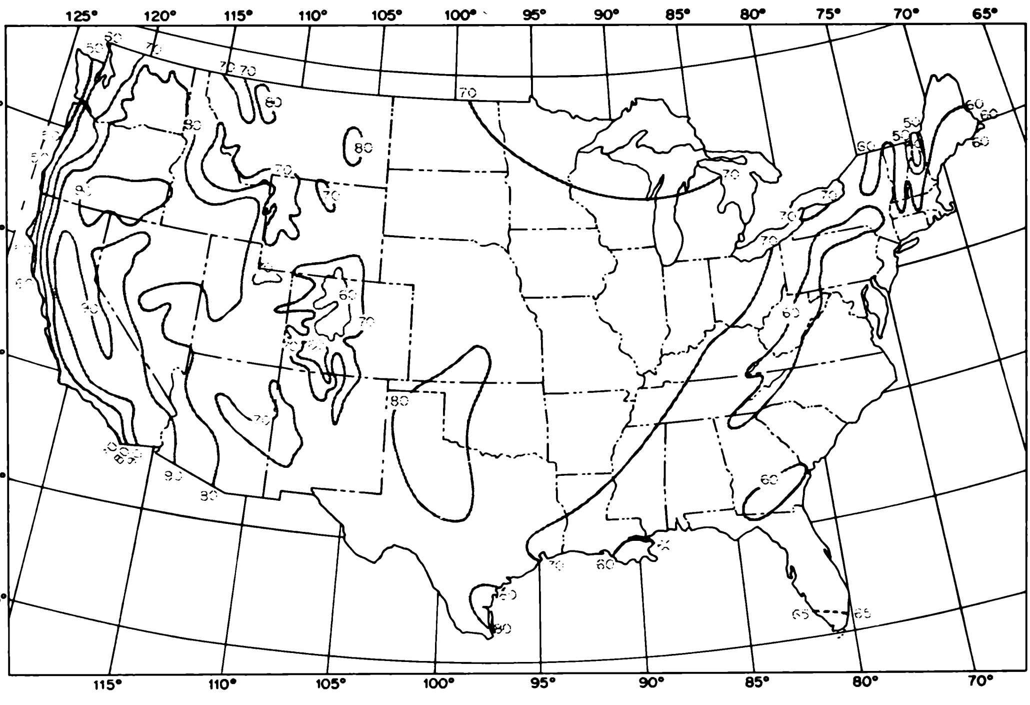


Figure 15.4
Mean percentage of possible sunshine in January in the contiguous United States. (NOAA, Environmental Data Service)

Figure 15.5
Mean percentage of possible sunshine in July in the contiguous United States. (NOAA, Environmental Data Service)



Tourist centers and resort areas are always proud to advertise whatever favorable sunshine data they can glean from climatic records. It is true that there are some large regional variations in duration of sunshine. Moreover, the intensity of sunlight is of primary significance in its effect on the body. At high latitudes and in the morning and evening hours, the low angle of the sun's rays makes sunlight much less potent. Smog and haze also detract from its power. At high altitudes, the air is much clearer so that solar radiation is more intense. Possible daily sunshine decreases with latitude in winter and increases in summer; the actual average duration varies with the climate, specifically with cloudiness. Figures 15.4 and 15.5 show the distribution of percentage of possible sunshine in the contiguous United States in winter and summer, respectively. It is evident that the drier climates are generally favored with more sunshine, but the claims of Florida are fairly well substantiated, especially in winter.

AIR POLLUTION AND HEALTH

The kinds of atmospheric pollution that may affect human health and comfort include gases (and their odors), liquids, inorganic solids, pollens and organic dusts, heat, light, noise, turbulence, and radioactivity. Their great variety requires a broad definition of atmospheric pollution: an undesirable concentration of any phenomenon in the air. Obviously pollution is a matter of degree as well as kind. Much less is known about the cumulative and long-term influence of moderate pollution on health than about specific disasters arising from abnormal concentrations, yet the former are equally important to human welfare. Moreover, the effects often are difficult to distinguish from other ailments. The incidence of harm from pollution has been highest among infants, the aged, and persons who were already ill.

Chemical effluents and solid particles are responsible for irritation of eyes and mucous membranes, intensification (if not the cause) of respiratory disorders, and pathological conditions leading to death. The more common chemical pollutants are sulphur dioxide, carbon monoxide, nitrogen oxides, ozone, and various hydrocarbons. Many airborne pollutants ordinarily occur in minute quantities, but some (for example, lead and arsenic) tend to accumulate in the body.

Although air pollution has no strict climatic boundaries, hazardous concentrations are more likely to develop in stable air. From a review of the physical causes of atmospheric stability it is evident that the conditions for a potential hazard are present in topographic depressions that are subject to air drainage and radiation inversions and in regions of stagnating anticyclones. These have been the circumstances of major air pollution disasters since the beginning of the industrial revolution. In December, 1930, 63 persons died as a result of intense industrial pollution in the Meuse Valley of Belgium. Several others suffered respiratory ailments, and many animals died. At Donora, Pennsylvania, in the Monongahela Valley, 21 people died during a pollution episode in late October, 1948. Thousands experienced respiratory and related disorders. The worst air pollution disaster on record occurred in London during the period December 5 to December 9, 1952, when a subsiding continental polar air mass trapped a shallow layer of smoke over the Thames Valley. In the Greater London area nearly 4,000 deaths were attributed primarily to sulphur dioxide effluents of industrial origin.

Events such as the foregoing have awakened government agencies to the need for control measures as well as for an improved understanding of the atmospheric conditions that enhance pollution. Urban and regional planning increasingly recognizes the topographic and climatic aspects of air pollution. In the United States the federal Air Quality Act of 1967 provided, among other things, for the designation of air quality control regions as a geographic framework for a systematic approach to air pollution problems. A Clean Air Act took effect in Britain in 1956. To facilitate public health warnings and the regulation of contaminating sources, the relative concentrations of one or more pollutants are used increasingly as bases for an *air pollution index*. Combined with weather forecasts and data on local pollution climatology, the index is an aid in averting disasters.

Natural aeroallergens are a form of pollution having close relations with weather and climate. Many plants have a narrow range of temperature and moisture requirements for the production, release, and dispersal of allergenic substances. Hay fever, asthma, and certain other respiratory ailments are initiated or aggravated as pollens and organic dusts are spread, often for great distances, by the wind. The end of the growing season for plant sources normally brings a halt to the "hay fever season."

As a type of natural pollution noise may be generated directly by wind and precipitation. Certain man-caused noises originate more often in one kind of weather than another. The roar of power mowers in the suburbs has a close correlation with sunny weekends. Weather conditions also influence the propagation of noise. Unstable air and turbulence in strong winds dissipate sound rapidly. Stable air associated with surface inversions of temperature and light, steady winds tend to concentrate sound near the ground, where it projects for great distances.

CLIMATE AND DISEASE

Many factors, such as cleanliness, nutrition, physical activity, and social contacts, act together in determining the incidence, severity, and spread of a disease. Climate is another factor, one which varies in importance according to the disease in question and the physiological and cultural characteristics of potential or actual victims. The climatic relations of all the common diseases cannot be discussed here; rather the aim will be to outline briefly the broad influences which weather and climate may exert on disease. The relations are extremely complex. Much cooperation in research among specialists in branches of climatology and of medicine is necessary to determine the role of climate in causing, modifying, or facilitating recovery from a specific disease.

There are two basic aspects of climatic influence on disease: the relationship of climatic factors to disease organisms or their carriers and the effects of weather and climate on the body's resistance. Many diseases are associated primarily with certain climates or with a season because of the temperature, moisture, and other requirements of the microscopic organisms that cause them. A number of parasites which attack humans are confined to the tropics and subtropics, where they find suitable conditions of warmth and moisture. Scarlet fever is virtually unknown in the tropics, whereas leprosy flourishes there. Some diseases depend upon intermediate carriers and are restricted to environments favorable to those carriers. Yellow fever and malaria, for example, are spread by certain species of mosquitoes that thrive in tropical climates. Rocky Mountain spotted fever occurs in the summer when the tick carriers are active.

Many diseases follow a distinct seasonal pattern. (See Figure 15.6.) Pneumonia and influenza are common seasonal diseases of the mid-latitudes; their greater incidence in winter is probably due to the lowered resistance in the upper respiratory tract at that season. Measles and scarlet fever cases are most numerous in spring. Infectious diseases occurring chiefly in winter and spring are much more widespread among the population than those that have their maximum frequency in summer and autumn.

Few diseases are caused directly by climate. A given combination of climatic elements may modify the metabolic rate, respiration, circulation, and the mental outlook of the individual so as to either strengthen or weaken his resistance to disease. Chilling, for example, lowers body resistance to most illnesses. Even in the tropics, sudden decreases in air temperature may be followed by outbreaks of sickness. Persons who fail to adjust their rate of physical activity in high or low temperatures often suffer some degree of exhaustion and a predisposition to disease. The

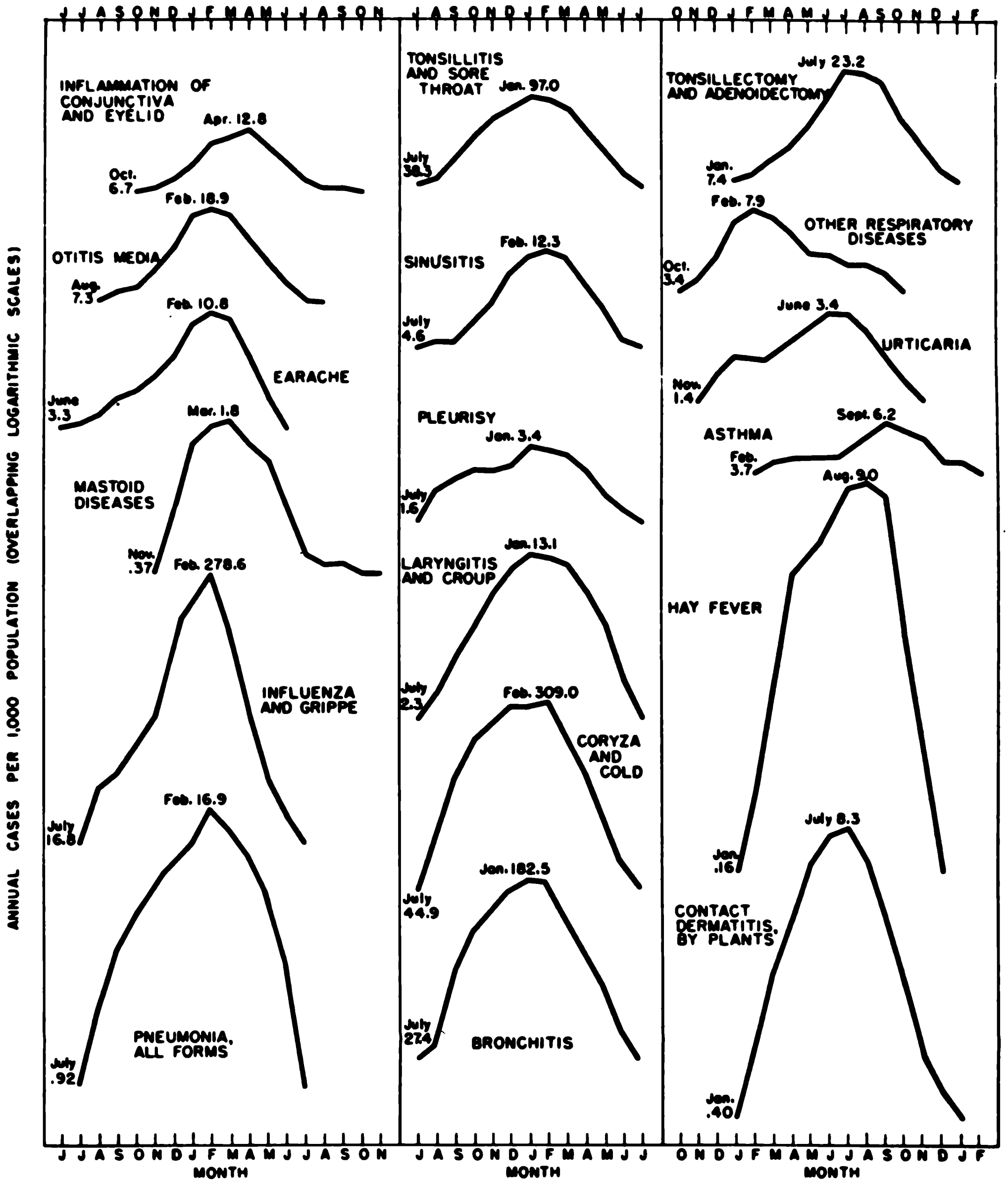


Figure 15.6
Relative seasonal variation in incidence of respiratory diseases and allergies in the United States. (Data from U.S. Department of Health, Education, and Welfare)

strain imposed by low barometric pressures at high altitudes also enhances development of a number of diseases. As we have seen, weather conditions can effect a concentration of pollutants and thereby increase their potential influences on health.

Favorable atmospheric conditions can assist the body in warding off disease and in promoting recovery if the disease is contracted. Fresh air, sunshine, mild temperatures, and moderate relative humidity all have therapeutic values. Fresh air and sunlight have long been recognized in the treatment of tuberculosis. Rickets and certain skin diseases respond to sunshine. The change of climate often prescribed for various kinds of illness is beneficial only if it is accompanied by rest, improved mental outlook, proper medical care, and good food. Moving to a location with a quite different climate may introduce problems that offset any actual curative powers of the new climate. Functional ailments which are obviously psychosomatic often are alleviated by moving to a different climate if emotional strain on the patient is thereby reduced. In such cases the change of scenery and social environment may be far more important than the change in atmospheric conditions.

ACCLIMATIZATION

Acclimatization is the process by which man (and animals other than man) becomes adapted to an unfamiliar set of climatic conditions. In the broad, popular sense it implies adjustment to all phases of a new physical and cultural environment, and indeed it is difficult to distinguish the purely climatic phenomena from other factors in strange surroundings. The chief problems in adjustment to a new climate frequently are not climatic at all but arise from homesickness, boredom, or social incompatibility.

In the narrower sense of physiological climatology, acclimatization entails actual changes in the human body brought about by climatic influences. It connotes a decrease in physiological strain as the body continues to be exposed to the new conditions. Temporary adjustments are made to daily and seasonal weather changes. But, when a person moves to a different climate, a more permanent adaptation gradually takes place. As would be expected from the foregoing sections, temperature is the element of greatest significance in acclimatization. Table 15.1 summarizes major physiological responses to high and low temperatures. In a hot climate the cutaneous blood vessels dilate so that more blood will be exposed to the low cooling power of the air and thereby maintain normal body heat. The blood supply is gradually increased in volume in order to meet the greater capacity of the capillary vessels, and it undergoes chemical changes. A larger number of sweat glands become active and there is a concomitant increase in thirst. Oxygen consumption declines. During the period of adjustment there may be physical discomfort, lack of energy, and an indifferent appetite. Persons in poor health may find their condition worsened by the strain. They are well advised to make the shift between two widely differing climates in gradual stages.

TABLE 15.1
Responses of the Human Body to Thermal Stress*

At low temperatures	*At high temperatures*
TEMPERATURE REGULATION	
Constriction of skin blood vessels	Dilation of skin blood vessels
Concentration of blood	Dilution of blood
Flection to reduce surface exposure	Extension to increase exposure
Increased muscle tone	Decreased muscle tone
Shivering	Sweating
Inclination to increase activity	Inclination to decrease activity
CONSEQUENT DISTURBANCES	
Increased urine volume	Decreased urine volume
	Mobilization of tissue fluid
	Thirst and dehydration
Danger of inadequate blood supply to exposed parts; frostbite	Reduced blood supply to brain; dizziness; nausea; fainting
	Reduced chloride balance; heat cramps
Discomfort leading to neuroses	Discomfort leading to neuroses
Increased appetite	Decreased appetite
FAILURE OF REGULATION	
Falling body temperature	Rising body temperature
Drowsiness	Impaired heat regulating center
Cessation of heartbeat and respiration	Failure of nervous regulation; cessation of breathing

*Adapted from *Arid Zone Research, X — Climatology, Reviews of Research,* Propioclimates of Man and Domestic Animals. Reproduced by permission of UNESCO. © UNESCO, 1958.

In some cases a satisfactory stage of acclimatization may never be achieved.

Acclimatization is not so clearly defined in a cold climate as in a warm one. Residents of cold climates cannot withstand continuous exposure to extreme cold but they are less affected than nonacclimatized persons. Repeated contact with cold air leads to a decrease in the flow of blood through constricted capillaries near the skin. Blood viscosity and the number of circulating white cells increase; the liver enlarges; appetite and oxygen consumption both increase in a natural attempt to provide more body heat. Sweating does not occur so readily and tends to be confined to certain skin areas, especially the palms, soles of the feet, and the axillae.

At high altitudes, adjustment must be made to both the lower temperatures and low pressure. Increased pulse and respiratory rates are the principal modifications which occur to offset the reduced oxygen supply.

People who live at high elevations develop greater lung capacity and larger chest cavities than residents of lowlands.

Although the human body is capable of withstanding remarkable extremes of climate, it functions best under circumstances where there is a reasonable degree of comfort and freedom from strain on vital organs. In general, children and healthy persons are more adaptable to a change of climate than the aged or infirm. Fortunately, it is possible to lighten the strain of acclimatization by proper diet, clothing and housing, or by control of physical activity. Many questions concerning these and other factors which modify climatic influences still remain to be answered by research in physiological climatology and related fields.

PERCEPTION OF CLIMATE

In addition to the daily adjustments people make to the changing weather they face decisions involving climate when planning a vacation or when they contemplate a change of residence for any reason, be it improved economic or social status, retirement, health, or climate itself. It has become increasingly practicable for people to move to places they perceive to have a "good" climate. Fortunately climate and other factors do not lead to the same choice for everyone. Unquestionably the dry summer subtropics have a high rating throughout the world. Arid and semiarid climates having moderate temperatures are also attractive to many, and a climate characterized by warm winters is usually preferred to a cold-winter type, hence the annual migration between the midlatitudes and the subtropics in Europe, North America, and Australia. For many people moderate to warm temperatures and abundant sunshine are the cardinal attributes of a pleasant climate. The cheering effect of a sunny day following a surfeit of gray skies is no mere illusion. Yet long periods of clear skies and glaring sunshine can induce ennui. Change, too, has its advocates if the extremes do not become too great. It may never be possible to identify, let alone quantify, all the variables that underlie perception of climate. The following quotation illustrates in brilliant literary fashion the subjective nature of the problem:

> I've lived in good climate, and it bores the hell out of me. I like weather rather than climate. In Cuernavaca, Mexico, where I once lived, and where the climate is as near to perfect as is conceivable, I have found that when people leave there they usually go to Alaska. I'd like to see how long an Aroostook County man can stand Florida. The trouble is that with his savings moved and invested there, he can't very well go back. His dice are rolled and can't be picked up again. But I wonder if a down-Easter, sitting on a nylon-and-aluminum chair out on a change-

> lessly green lawn slapping mosquitoes in the evening of a Florida October—I do wonder if the stab of memory doesn't strike him high in the stomach just below the ribs where it hurts. And in the humid ever-summer I dare his picturing mind not to go back to the shout of color, to the clean rasp of frosty air, to the smell of pine burning and the caressing warmth of kitchens. For how can one know color in perpetual green, and what good is warmth without cold to give it sweetness?*

Fundamentally, human perception of climate embraces the criteria of cause, magnitude, time and duration, spatial arrangement, uncertainty, and resultant effects. Rarely do different individuals apply these criteria in the same way. For example, at one extreme primitive man may see a climatic event as a supernatural act; at another the scientist seeks to define it in mathematical terms as a system of physical processes. Each perceives the effects in the light of his personal experience and beliefs. Climatic hazards such as droughts, floods, severe storms, or heat waves thus evoke a broad spectrum of cultural perspectives that vary not only among individuals but also with time. Whereas popular accounts of past climatic disasters may sometimes exaggerate the facts, it is common for people to minimize the threat of predicted storms and to resist procedures that might help them avert a catastrophe.

Beliefs relating to climatic change are a particularly intriguing aspect of the perception of climate. The layman's concern about a possibly changing climate is accentuated by extreme weather events more often than by gradual trends. Even where the average temperature may have risen as much as 2° or 3C° the change would go unnoticed by most people, unless it is manifested in extremes. The increased use of cooling systems in homes and buildings is clearly more the result of their greater availability at moderate cost than any real effect of a warming climate. Air conditioning, transportation, eating habits, clothing, agricultural methods, and many other attributes of culture have changed to such a degree that we cannot rely on personal evaluations of climatic change. It is entirely possible that the knee-deep snow of Grandfather's youth impressed him more than the same depth would today because he walked to school and perhaps helped with the farm chores morning and night regardless of the weather. Throughout the course of human history technology has generated a changing attitude toward resources and the natural environment, including climate.

---

* From TRAVELS WITH CHARLEY *In Search of America* by John Steinbeck. Copyright © 1961, 1962 by Curtis Publishing Co., Inc. Copyright © 1962 by John Steinbeck. Reprinted by permission of The Viking Press, Inc., and William Heinemann, Ltd., British publishers.

Additional Readings

AULICIEMS, ANDRIS, *The Atmospheric Environment: A Study of Comfort and Performance.* Toronto: University of Toronto Press, 1972.

CASTELLANI, ALDO, *Climate and Acclimatization,* 2nd ed. London: J. Bale Sons and Curnow, 1938.

ELLIS, F. P., "Mortality from Heat Illness and Heat-Aggravated Illness in the United States," *Environ. Res.,* 5, 1 (1972), 1–58.

FALCONER, RAYMOND, "Windchill, A Useful Wintertime Weather Variable," *Weatherwise,* 21, 6 (1968), 227–29, 255.

GOLDSMITH, JOHN R., AND STEPHEN A. LANDAW, "Carbon Monoxide and Human Health," *Science,* 162, 3860 (1968), 1352–59.

KERSLAKE, D. MCK., *The Stress of Hot Environments.* New York: Cambridge University Press, 1972.

LANDSBERG, H. E., *Weather and Health.* Garden City, N.Y.: Doubleday & Company, Inc., 1969.

LEE, DOUGLAS H. K., AND HOYT LEMONS, "Clothing for Global Man," *Geog. Rev.,* 39, 2 (1949), 181–213.

LICHT, SIDNEY, ed., *Medical Climatology.* New Haven, Conn.: Elizabeth Licht, Publisher, 1964.

"Physiological Principles; Comfort and Health," Chap. 7 in *ASHRAE Handbook of Fundamentals.* New York: American Society of Heating, Refrigerating, and Air-Conditioning Engineers, Inc., 1972.

RENBOURN, E. T., "Tropical Clothing, A Physiological Appreciation," in *Biometeorology,* ed. S. W. Tromp. London: Pergamon Press, 1962, pp. 91–98.

SARGENT, FREDERICK, II, AND SOLCO W. TROMP, eds., *A Survey of Human Biometeorology,* Tech. Note No. 65. Geneva: WMO, 1964.

STEPHENSON, P. M., "An Index of Comfort for Singapore," *Met. Mag.,* 92, 1096 (1963), 338–45.

THOM, E. C., "The Discomfort Index," *Weatherwise,* 12, 2 (1959), 57–60.

TROMP, S. W., et al., *Medical Biometeorology.* Amsterdam: Elsevier Publishing Company, 1963.

WINSLOW, C. E. A., AND L. P. HERRINGTON, *Temperature and Human Life.* Princeton: Princeton University Press, 1949.

chapter sixteen

Climatic Change

In the preceding chapters we have examined some of the many relationships between climate and human activities on earth. For the most part we have been concerned with the contemporary climatic pattern and its effects on life. We take for granted the weather changes that collectively make up climate; in fact, the only constant feature of weather is change. No one denies the cyclical nature of diurnal and seasonal weather fluctuations, and cyclonic storms form and dissipate with a certain regularity. But several kinds of evidence point to variations in climate as well. The implications of a rapid secular change in climate are so complex and so awful as to defy the imagination. Agriculture, manufacturing, commerce, and all other human endeavors would need to adjust dramatically if the world's climates were to change greatly during a generation. Gradual upward trends in mean temperature of only a few degrees could alter the global biotic pattern and even eliminate certain forms of life. Some people become alarmed at the prospect of polar warming, for the polar icecaps contain enough water to raise sea level by 50 to 80 m. Studies thus far indicate that increased temperatures on the order of those experienced during the last century would have to prevail for 10,000 years to melt all of the polar ice.

Climate as a summation of atmospheric conditions over an infinite time span does not change. Climatic change has no meaning unless we compare specific periods and analyze trends. For the purpose of examining past climatic changes periods of time and the associated evidence fall into three broad categories:

1. Periods on the order of millions of years—*paleoclimates.*
2. Periods in the last several thousand years, whose climates may be interpreted in terms of natural phenomena during recent geological epochs.
3. Historic time, the latter part incorporating climatic records.

PALEOCLIMATOLOGY

Although abundant, and sometimes contradictory, evidence indicates marked changes in climate since the creation of the earth, the data become increasingly satisfactory through the eons. We know little about earth climates prior to the beginning of the Cambrian period nearly 600 million years ago, when life forms began to develop rapidly and the free oxygen content of the atmosphere increased. Fossilized remains of plant and animal life in various sedimentary deposits are the primary clues to inferences about the duration and geographical extent of temperature and moisture conditions in early geological history. They are the basic data of paleoclimatology, which borrows its methods from paleontology. Inasmuch as fossil evidences of life span the entire chronology of paleoclimatic studies, it follow that the range of climatic fluctuations must fall within the limits for sea life and probably within the limits for plant life on land as well. Thus, sustained global temperatures such as 100°C or −50°C are highly unlikely features of paleoclimates.

A reconstructed calendar of paleoclimates shows a 150-million-year period of high temperatures through the Cambrian, Ordovician, and Silurian, that is, between 500 and 350 million years ago. (See Figure 16.1.) Since then the general trend of temperature has been downward, although there have been significantly large variations. During most of the time since the Cambrian the polar areas of both hemispheres were ice-free, but near the end of the Paleozoic glaciation was widespread for 30 to 50 million years over Gondwanaland, a large southern continent comprising present-day Australia, Antarctica, South and Central Africa, and South America. Inferred paleotemperatures indicate surface values on the order of 8° to 10°C near the poles and 25° to 30°C in the tropics during the Tertiary period, when the subtropic highs gradually shifted from 50° or 60° latitude toward their present positions. The lowest Northern Hemisphere temperatures since the Cambrian prevailed dur-

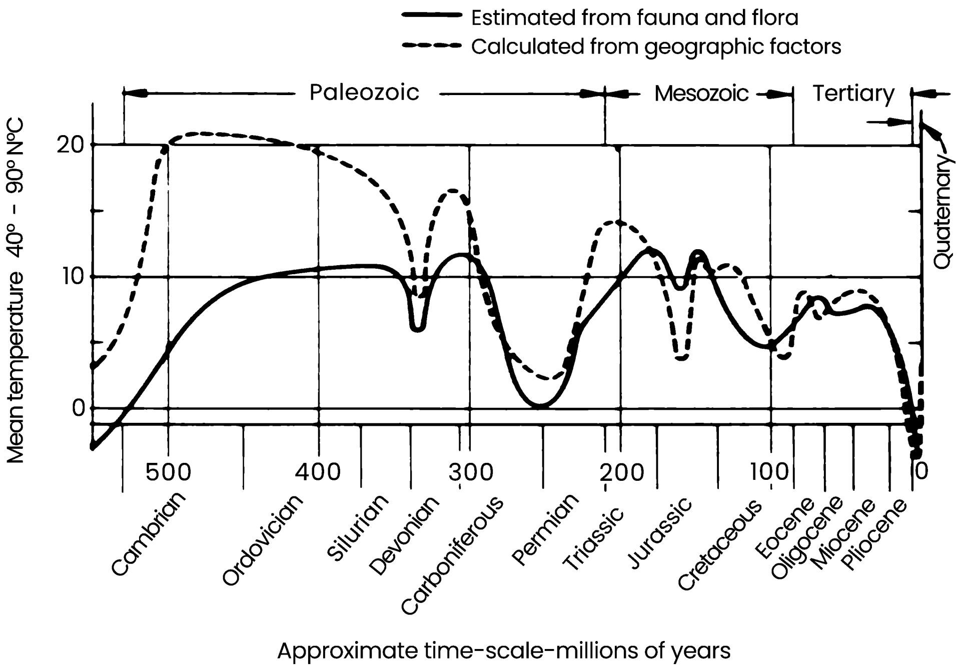


Figure 16.1
Variations of temperature during geologic time. (From an article by C. E. P. Brooks, "Geologic and Historical Aspects of Climatic Change," *Compendium of Meteorology*, 1951, published by the American Meteorological Society through support extended by the Geophysics Research Division. Air Force Cambridge Research Center, Air Research and Development Command. Reproduced by permission)

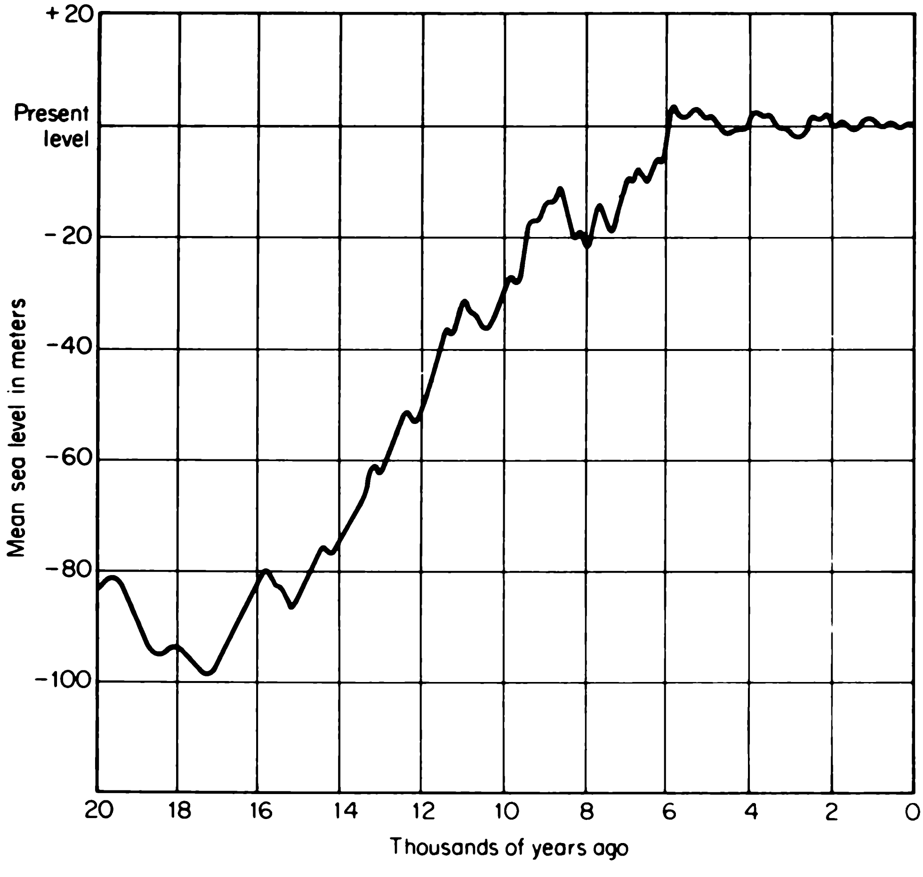
ing the glacial ages of the Pleistocene epoch, beginning more than a million years ago. Cold interludes of extensive continental glaciation then alternated with warmer interglacial periods. The last major ice advance reached its maximum about 18,000 years ago and was halted by rapid melting and recession that continued until about 6,000 years ago. Concurrently mean sea level rose aproximately 10 m. (See Figure 16.2)

GEOCHRONOLOGY OF CLIMATES

Knowledge of the sequence of past climatic events is basic to all studies of climatic change, but a chronology is essential for correlating the evidence derived from related phenomena and for worldwide comparisons. Dating of paleoclimates is accomplished primarily by reference to rocks, especially those in sedimentary strata. Several methods that depend on the rates of decay of radioactive isotopes assist the dating of geological formations, including deep-sea deposits. An oxygen-isotope method, discovered by Harold C. Urey, has been applied to fossil shells to estimate and date sea temperatures extending back more than 400 million years. Techniques for dating organic materials are especially useful for correlating the evidences from life forms and geology. One such technique is the carbon-14 method developed by Willard F. Libby. It is applicable to plant and animal remains and even to carbon dioxide in air and water but has limited chronological range. Paleosoils, ancient peat beds, residues from the evaporation of water bodies (*evaporites*), and cores from the depths of ice caps and the sea bottom are among the types of indirect climatic evidence that can be dated by radioisotope analysis.

One of the most detailed climatic records for the past few thousand years lies in *varves*, the annual layers of silt and clay deposited on the bottoms of lakes and ponds that are subject to freezing in winter and thawing in summer. (See Figure 16.3.) The only material being deposited in a frozen lake is the fine clay which is held in suspension; the surface ice prevents other materials from entering. When thawing begins, fresh

Figure 16.2
Changing sea level during the past 20,000 years. (After Fairbridge)



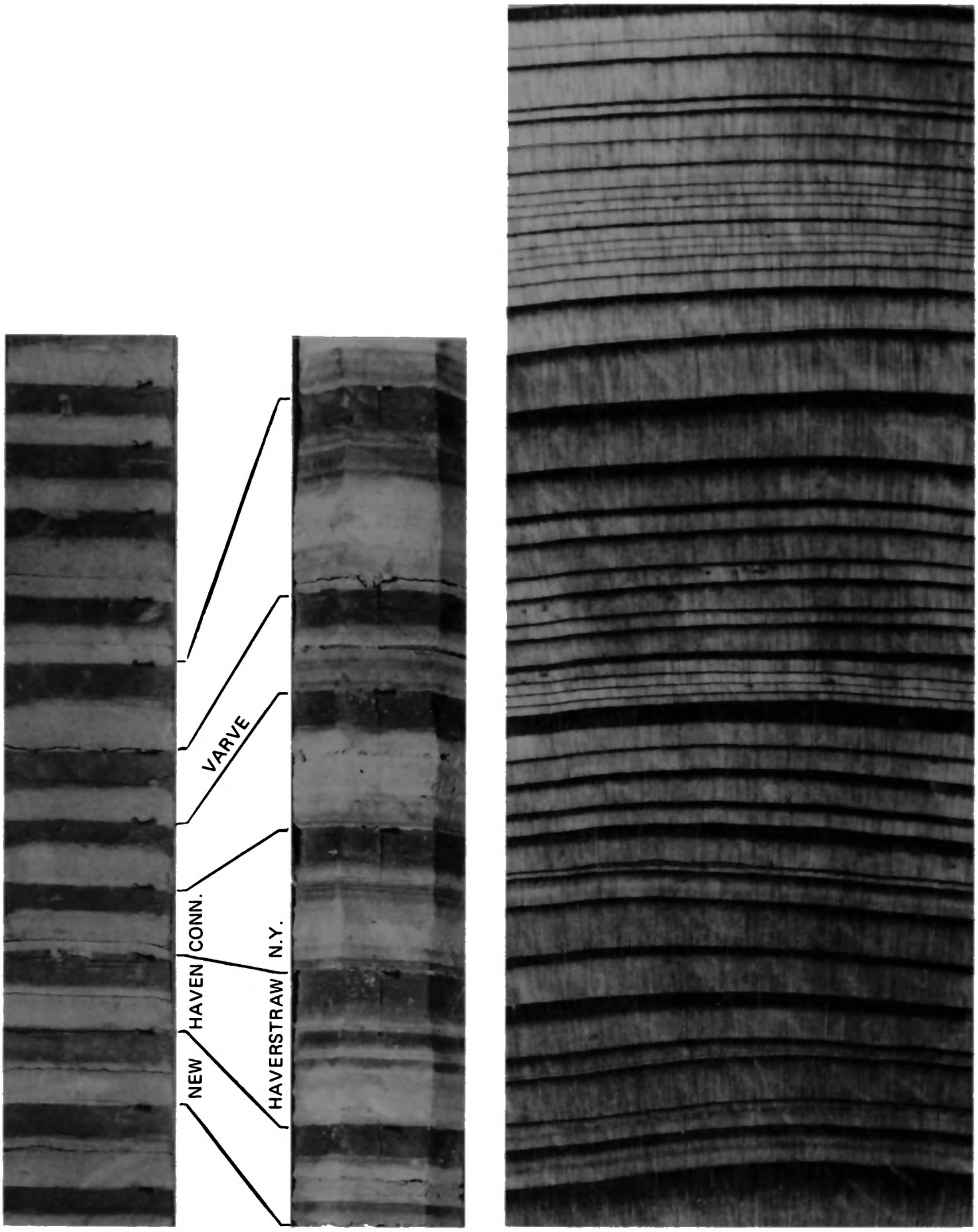


Figure 16.3
Examples of geochronological evidence. Varve correlations on the left are from sites near New Haven, Connecticut, and Haverstraw, New York. The tree-ring cross section on the right indicates effect of rainfall variations on growth of a Douglas fir in northeastern Arizona. (Varve photo courtesy The American Museum of Natural History; tree-ring photo courtesy Laboratory of Tree-Ring Research, University of Arizona)

water and new sediments are introduced. Presumably because of climatic fluctuations, no two successive years have the same thickness of deposits. Thus, parallel dating of lake beds in the same region is possible through correlation of the thickness of the varves. Varves cannot be formed under glaciers because of the necessity for annual freezing and thawing, and, for this reason, a comparison of lake beds in once-glaciated areas enables geochronologists to determine accurately the dates of recession of glaciers. About 13,700 years are represented in the varve records of Scandinavia, where the oldest varves are in the south and the newest in the north, thus making it relatively easy to trace the retreat of the last ice sheet.

A second technique of geochronology is based on the annual increment of tree growth and is termed *dendrochronology*. (See Figure 16.3.) Records in tree rings go back about 3,000 years in living trees, but study of fossil trees and ruins may extend this back another 1,000 years or more. A close relationship between annual rainfall and the growth rings of trees under climatic stress was demonstrated by A. E. Douglass in the southwestern United States, where fluctuations in precipitation have been fairly well charted. Studies by Douglass and others at the University of Arizona show four interesting features of climate over the past 850 years in the interior plains and plateaus of the United States and Canada. Almost the entire thirteenth century was dry, and the fourteenth was stormy. The last quarter of the sixteenth century was the driest period in the last 650 years. About 1670 there developed a remarkable change from long-term irregular fluctuations to shorter ones of only 20 or 25 years. The drought years just before 1900 and in the 1930s are examples of the latter cycles.

Dendrochronology has also been employed to date the comparatively recent advance and retreat of glaciers. Examination of cross sections of trees partly pushed over by glacier ice along the margins of its maximum advance but left alive in a tilted position reveals the exact year of the maximum advance of the glacier. While the tree was erect growth rings were concentric, but after tilting due to ice pressure the rings lost their symmetry. Narrow growth rings give evidence of periods when ice was close enough to a tree which remained erect to affect the temperature.

One of the most fruitful fields of investigation into recent climatic chances is glaciology. Consistent trends in precipitation and temperature are reflected in the advance and retreat of glaciers. Most of the direct study of glacier fluctuations and climatic relationships has been carried out in the twentieth century. The foremost student of glaciation and climatic variation, the Swedish glaciologist Hans Ahlmann, first correlated measurements of ice ablation (reduction of glacier volume by evaporation, sublimation, and melting) and simultaneous weather observations. The extent of glaciers is affected by annual snowfall, duration and degree of temperatures above the melting point, and the amounts of incoming and outgoing radiation. These factors are affected, in turn, by cloudiness, wind speed, and humidity. The shape of the basin or valley containing the glacier also influences its movement; some glaciers cannot advance until they have increased in volume sufficiently to surmount topographic barriers.

Glacier observations by Ahlmann and his followers have shown that the climate of the arctic and subarctic became warmer in the first part of the twentieth century, the major change being in the higher latitudes. This trend may have undergone a reversal, however.

When augmented by radioisotope dating techniques studies of existing glaciers and ice fields can also reveal data on climates of the distant past. The temperature at the time of sublimation determines the ratio of oxygen isotopes to H_2O molecules in crystals of snow or ice; this provides a basis for calculation of average air temperatures from polar ice samples. Application of the oxygen-isotope method of paleotemperature analysis to deep cores from the ice caps of Greenland and Antarctica has yielded temperature estimates extending back 100,000 years. One antarctic core, for example, indicates a relatively rapid temperature decrease on the order of 5C° about 60,000 years ago, followed by slow warming and then another cooling period beginning about 30,000 years ago.

Another technique of geochronology uses *pedogenic* criteria, that is, data obtained through study of the development of soils. Examination of old soils which have been buried in river flood plains, along fluctuating lake shores, or under windblown deposits furnishes some evidence of the nature of past climates, for climate is the most important factor in soil formation. A generalized method of climatic dating is possible through analysis of peat bogs. The most effective analysis has been accomplished by means of pollen studies. Under the microscope, pollen grains serve as identification of the plant associations that prevailed at various stages during the peat accumulation. The succession of plant types in peat shows that there have been several long climatic waves during the last 30,000 to 35,000 years. In the British Isles it has been possible to correlate peat studies with the cultural development of man.

Fossils of animals as well as plants aid in piecing together the history of climate over the past few thousand years. The bones of arctic mammals have been found far south of their present limits of distribution, and well-preserved remains of extinct animals occur in glacial ice in the arctic. The mastodon is perhaps the best-known example. In western Europe remains of desert and steppe mammals have been unearthed.

CLIMATE DURING RECORDED HISTORY

During the entire period of written and inferred human history the data of geochronology overlap archaeology, documentation, and instrumental records to serve as a double check on supposed climatic changes. It might be naïvely assumed that recorded human history would lend itself readily to analysis of past climates. On the contrary, documental climatology deals largely with manuscripts written for purposes other than climatic description, and interpretation is difficult. Problems arise from changes in the calendar and lack of continuity of historical records for a given location. Writers have had a tendency to mention droughts, severe storms, or other extremes and generally have neglected periods of relatively stable climatic conditions. Nevertheless, records of crop yields or crop failures, of floods, and of migrations of people furnish useful evidence of the possible influence of a changing climate. Correlation with the geochronological evidence yields the following pattern since the birth of Christ.

As would be expected, there are more data for Europe than for other parts of the world. During the first century of the Christian era, the pattern of precipitation in Europe and the Mediterranean resembled that of today. This was followed by a wetter period ending about A.D. 350. The fifth century was dry in Europe and probably also in North America, where many of the western lakes dried up completely. In the seventh century Europe was both warm and dry. There was heavy traffic over passes in the Alps that are now filled with ice. Tree rings show a dry period in the United States at this time, and the Nile floods were low. Europe was wetter again in the ninth century. During warm, dry conditions in the tenth and eleventh centuries wine production reached its peak in England. Vikings settled Greenland in 984 in a time of relatively favorable climate, only to abandon it about A.D. 1410. The first half of the thirteenth century was stormy in the North Sea and North Atlantic. A drought from A.D. 1276 to 1299 is believed to have driven the cliff dwellers out of Mesa Verde in Colorado. The fourteenth century was cold and snowy in northern Europe and northern North America. The Aztecs settled Mexico in 1325, when lakes were at higher levels than today. The so-called "Little Ice Age" extended from 1550 to about 1850. During that period European glaciers advanced farther than at any time since recession of continental ice. Arctic sea ice was also probably much more extensive than at present. Since the middle of the nineteenth century Northern Hemisphere glaciers have retreated to their sixteenth-century position, and sea level has risen by 5 to 8 cm.

RECENT CLIMATIC TRENDS

Finally, let us consider climatic changes in the very short period for which we have instrumental records. At the beginning of the nineteenth century there were five places in the eastern United States and twelve in Europe where satisfactory weather observations were being taken. Not until 1876 did the climatic record include any significant weather records taken at sea. Like all other forms of historical data relating to weather and climate, instrumental records fade away rapidly into the past, but instrumental climatology was fairly well established by 1850. One would expect that records of instrumental observations would bring an end to speculation on recent climatic changes. The statistician, treating the data objectively, should provide the facts on trends in temperature and precipitation, the two elements for which records are most abundant. But climatic records are not readily subjected to objective study. In the first place, weather observers are subject to human failings, and even small errors affect calculations that may involve equally small trends. The exposure and height above ground of instruments also materially affect results. Removal of a weather station to a new location practically destroys the value of its records for purposes of studying climatic change. But, even if a station remains in the same location for a century, the changes in vegetation, drainage, surrounding buildings, and atmospheric pollution are likely to produce a greater effect on the climatic record than any true climatic changes. Thus, very careful checking and comparison of climatic records is necessary to detect climatic fluctuations. In order to overcome some of the problems arising from changes in station location or in local environment the United States Environmental Data Service has established a *climatological benchmark network*. Chosen for their representative sites, continuity of past records, and promise of future permanency, these benchmark stations will provide basic data against which other climatic records can be evaluated.

Statistical analyses of temperature records in the United States show that, in the first half of the twentieth century, there was a general warming trend, but that the change was small. The average for many stations and for the year increased by less than 1C°, reaching a peak about 1940. Winter temperatures increased slightly more than those for summer months. However, the trend was not the same in all parts of the country. In the Pacific Northwest, northern Rockies, and the northern Great Plains, mean temperatures of the winter months decreased or showed little change.

World temperature trends were upward from about 1885 to 1940, the amount of rise being about 1C° in winter and less than 0.6C° for the year. The greatest increases were in winter over the arctic regions, where rises in excess of 3C° characterized the period 1917 to 1937. After 1940 the rate of increase slowed and then began a reversal. (See Figure 16.4.) Low-latitude regions experienced comparable trends.

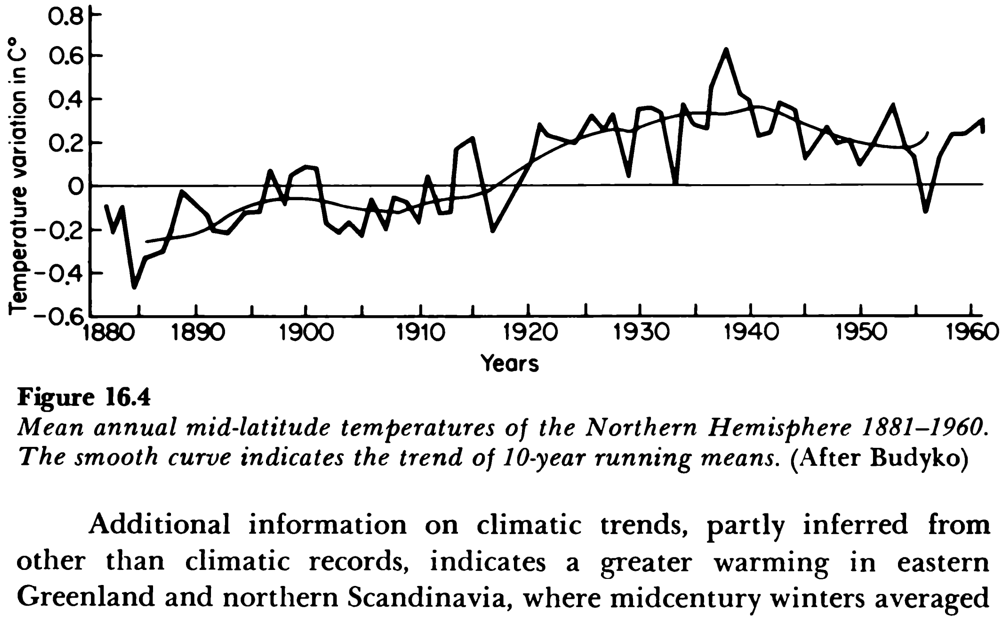


Figure 16.4
Mean annual mid-latitude temperatures of the Northern Hemisphere 1881–1960. The smooth curve indicates the trend of 10-year running means. (After Budyko)

Additional information on climatic trends, partly inferred from other than climatic records, indicates a greater warming in eastern Greenland and northern Scandinavia, where midcentury winters averaged 4° to 7C° higher than in 1900. This warming was expressed in the oceans also; codfish were found farther north in the Atlantic than ever before in history.

CLIMATIC CYCLES

Although it seems clear that there have been climatic fluctuations in historic times, there is much less certainty about the regularity of climatic cycles. Even the well-recognized daily and seasonal cycles vary in length by large percentages. Daily maximums of temperature, for example, rarely succeed one another at intervals of exactly 24 hours. At high latitudes the daily cycle takes on a character much different in summer than in winter. The annual cycle often has seasons that are early or late; January and July are by no means always the extreme months in middle and high latitudes. The reversal of the monsoon in southern and eastern Asia is generally regarded as an annual phenomenon, but it occurs at widely varying dates.

Statistical analyses have suggested a great number of supposed cycles, none of which has a dependable regularity. Most are barely distinguishable from chance anomaliees and properly should be termed rhythms or quasi-periodicities. They range in length from one year to millions of years. The two-year cycle which commonly appears in series of climatic data reflects persistence in atmospheric conditions, the heat budget in particular. For example, heat stored by the atmosphere, oceans, and land in one year tends to support higher temperatures well into the next. Investigations of climatec cycles often are more productive when cause and effect are treated together. The relentless search for periodicities in solar emisions, sunspots, earth-sun relationships, composition of the atmosphere, and other phenomena that may influence climate will no doubt continue as long as there is even a remote hope of finding true cycles, which would be valuable aids to climateic forecasting.

THEORIES OF CLIMATIC CHANGE

Few natural phenomena have attracted the attention of so many scientific fields or elicited so many hypotheses as have climatic changes. In view of the fundamental role of radiation in the energy processes that produce weather and climate it is understandable that most theories have dealt with the possible effects of an imbalance in the earth's heat budget. One of the simplest and most persistent holds that the sun is a variable star and that changes in the kind and amount of energy emitted alter the solar constant. Increases in solar radiation would warm the atmosphere and account for such phenomena as melting of the continental glaciers. Sir George Simpson theorized, however, that moderately increased insolation would permit higher moisture content of the air, cause stronger meridional transfer of air, and increase precipitation in polar areas. Greater summer cloudiness would inhibit melting of the accumulated snow and ice. Conversely, diminished insolation would weaken the general atmospheric circulation and thereby reduce precipitation at high latitudes. Thus, paradoxically, a lowering of the mean atmospheric temperature might cause a recession of ice sheets, whereas a temperature increase would cause them to advance.

Theories based on a changing sun merge with those concerning sunspot cycles. No completely acceptable theory has been developed to explain how solar variations are translated into climatic fluctuations, if indeed they are. Examination of the number of sunspots during more than two centuries reveals a sunspot cycle of about 11.3 years, but the period has been as short as 9 years and as long as 16 (see Figure 16.5). Secondary cycles having lengths of 35 years, 80 years, and other periods have also been suggested. Cooler, wetter weather and greater storminess have usually accompanied an increase in sunspot numbers; warmer, drier conditions are associated with a decrease. Thus, greater solar activity seems to produce lower atmospheric temperatures. The explanation of this paradox may lie in the effects of variable insolation on the general circulation of the atmosphere.

A possible link between solar variations and the general circulation has been elaborated by Hurd C. Willett. Short bursts of ultraviolet and corpuscular energy presumably are absorbed in the upper atmosphere, where resulting temperature variations are expressed as differences in pressure and in the "normal" pattern of tropospheric circulation. According to Willett, the vigorous emissions of ultraviolet radiation have their greatest effect in low latitudes, leaving the polar areas relatively cold and producing a zonal pattern of circulation like that of the hypothetical planetary wind and pressure systems. Corpuscular radiation consists chiefly of charged protons, and its maximum emission comes with solar disturbances. Corpuscular bursts are diverted by the earth's magnetic field toward the magnetic poles and thus tend to heat the upper air of the polar regions more than the tropics, especially in winter. As a consequence, the zonal circulation is disrupted and there is a

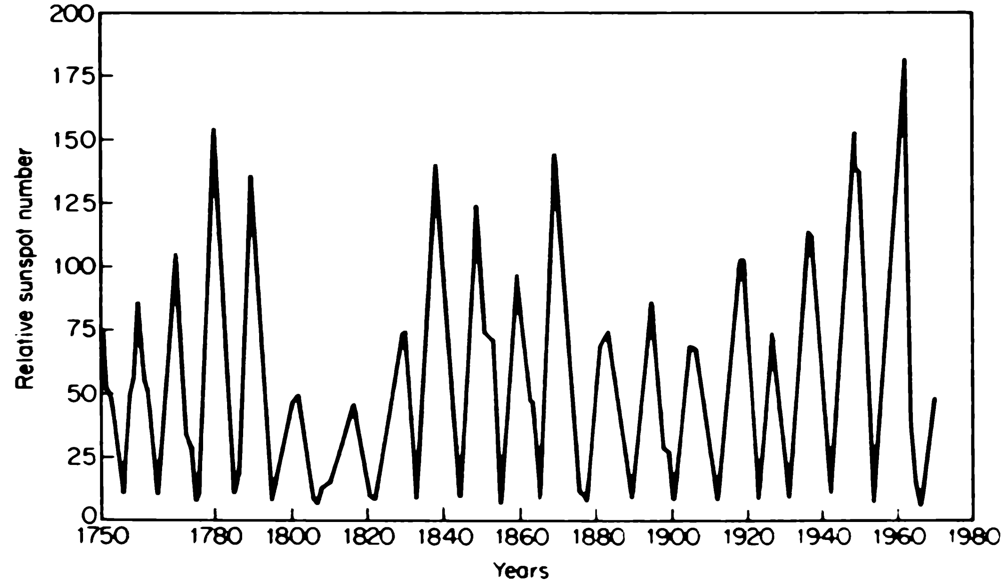


Figure 16.5
Variation of mean sunspot activity 1750–1970. (After Pomerantz and Swiss Federal Observatory, Zurich)

greater latitudinal transfer of air with accompanying storminess and temperature extremes. This theory does not require an appreciable change in average atmospheric temperatures but rather a modification of the general circulation wherein zonal circulation is interspersed with periods of nonzonal air movement and weather extremes.

Assuming no fluctuations in solar output, there are other possible causes of variable insolation. Astronomical theories consider four principal effects:

1. Changes in the angle which the earth's axis makes with the plane of the ecliptic. This would affect the seasons, temperature distribution, and the general circulation.
2. Changes in the eccentricity of the earth's orbit. Resulting variations in the mean distance from earth to sun could affect average temperatures on the earth.
3. Precession of the equinoxes. This refers to the regular change in the time when the earth is at a given distance from the sun. At present, the earth is closest to the sun in the Northern Hemisphere winter (about January 3rd). About 10,500 years ago the Northern Hemisphere winter came at a time of the year when the earth was farthest from the sun. Other things being equal, winters should have been colder, and summers hotter, than they are now. In the Southern Hemisphere the reverse applies.
4. Shifting of the earth on its polar axis. The supposition is that the geographic poles migrated to their present locations from what are now lower latitudes, there being a corresponding shift of the attendant climatic conditions.

Having reached the outer limits of the atmosphere, insolation is affected by the processes of absorption, reflection, and scattering. A prominent theory suggests that volcanic ash has increased the albedo of the atmosphere from time to time, and thus has reduced insolation at the earth's surface. Ash layers in antarctic ice show a period of intense volcanic activity from about 30,000 to 17,000 years ago, during which temperatures cooled by about 3C°. In the modern era the eruption of Mount Tambora on the Indonesian Island of Sumbawa in 1815 ejected an estimated 150 km^3 of ash into the atmosphere. The following year was known as "the year without a summer" in the United States and Europe. Explosions of ash from Krakatoa in the Strait of Sunda in 1883 produced red sunsets for many months and may have contributed to the severe winters that followed in the Northern Hemisphere. Massive eruptions of other volcanoes in the twentieth century (Katmai, 1912; Agung, 1963; Taal, 1965; Mayon and Fernandina, 1968) forced ash into the stratosphere. Solid particles resulting from wind erosion, burning of vegetation, and the pollution by cities and industries also contribute to atmospheric dust. Insofar as their increase is the result of human activities we may consider man as a factor in climatic change. (See Chapter 17.)

Closely allied to the dust theories are others that postulate variations in the amounts of atmospheric gases, especially those which absorb radiation selectively. Carbon dioxide and water vapor tend to create a so-called greenhouse effect by transmitting short-wave radiation but subsequently absorbing part of the long-wave terrestrial radiation. An increase in carbon dioxide is thought to produce slightly higher surface temperatures, and it has been proposed as a cause of the warming trend from about 1885 to 1940. (A possible explanation of the apparent cooling since 1940 is the overriding effect of rapidly increasing atmospheric dust.) A doubling of atmospheric CO_2 could raise surface air temperatures by 2C°. Variations in the amount and the height of maximum concentration of ozone in the upper atmosphere might also affect air temperatures. Ozone absorbs ultraviolet radiation from the sun as well as a portion of infrared terrestrial radiation.

The influence of the earth's surface on the heat and moisture budgets suggests another category of theories. Continental drift during the past geologic periods would account for climatic changes of major proportions as land masses shifted in relation to one another and assumed different latitudinal positions. Structural warping of ocean basins might alter sea level and oceanic circulation, consequently affecting the transport of heat and moisture. Mounting evidence points to a correlation between climatic fluctuations and circulation systems in the oceans. But the relationship is reciprocal. Climatic changes may be reflected in water temperatures and ultimately in oceanic circulation, which in turn affects climate. It thus becomes difficult to separate cause from effect.

If vertical temperature lapse rates similar to those of the present prevailed in the past, changes in elevation owing to major crustal upheavals or erosion might have initiated glacial fluctuations and redistribution of vegetation, both of which are important clues to past climates. Changes in surface albedo are other possible causes of a revised heat budget. It has been calculated that an increase in the mean albedo of the earth (including the atmosphere) by an additional 1 percent of total insolation would lower average earth temperatures by about 2C°. Freezing of ocean surfaces and formation of extensive ice sheets would greatly increase the proportion of reflected radiation. Such hypotheses rapidly lead to circular argument. If no other factors were involved, cooling would appear to beget glaciation, which in turn increases the surface albedo, producing further cooling. Obviously, other factors have been at work in climatic change; a single phase of the heat budget should not be examined without reference to the others. Nor can changes in the circulation patterns of the atmosphere and oceans be regarded as ultimate causes of climatic change, for these motion systems are themselves subject to climatic effects.

FORECASTING CLIMATE

Explanations of past climates logically lead to forecasts of future climates. Past rhythmic changes offer one of the best possibilities, but thus far no climatic cycles longer than one year have been defined adequately to permit acceptable forecasts for several years in advance. Most students of climatic change agree that the earth is currently in an interglacial period, but few venture predictions of the dates of the next ice age. If the apparent trends of the past two centuries repeat themselves, the last half of the twentieth century should be cooler and wetter than the first half in the United States. As always, we should expect years in which the weather departs from the average. Forecasts of climate, like weather predictions, cannot be evaluated until the end of the forecast period, but data for checking theories of climatic change indirectly may be available sooner. For example, meteorological and solar observatories beyond the outer limits of the atmosphere provide an increasing store of information on the components of both solar and terrestrial radiation.

If we accept the theory concerning the effects of increasing carbon dioxide in the atmosphere, the overall trend in climate during the next several decades should be toward warmer temperatures. The theory does not rule out the possibility of cyclical variations superimposed on the long-term trend, however. If we assign major importance to atmospheric dust, the increasingly turbid air should bring a temperature decline. If "clean" forms of energy replace coal and oil within the near future, the rate of increase of both carbon dioxide and dust should be curtailed. Any forecast that depends on cultural factors must allow for changing technology, which is fully as difficult to predict as natural phenomena. Who can guarantee that volcanic eruptions, geomorphic processes, solar or astronomical effects, or some other suspected or unknown cause of climatic change will not produce even more marked climatic fluctuations in the next few centuries than have occurred in the past?

Additional Readings

Abbot, C. G., *Solar Variation and the Weather,* Smithsonian Misc. Coll., 146, 3. Washington: Smithsonian Institution, 1963.

Brooks, C. E. P., *Climate Through the Ages.* New York: McGraw-Hill Book Company, 1949.

CALLENDAR, G. S., "Temperature Fluctuations and Trends over the Earth," *Quart. Journ. Roy. Met. Soc.,* 87, 371 (1961), 1–12.

Changes of Climate, Arid Zone Research XX. Paris: UNESCO, 1963.

DZERDZEYEVSKIY, B. L., "General Circulation of the Atmosphere as an Essential Link in the System: Sun-Climatic Change," *Soviet Geog.,* 5, 3 (1964), 37–52.

EWING, MAURICE, AND WILLIAM L. DONN, "A Theory of Ice Ages, I," *Science,* 123, 3207 (1956), 1061–66.

———, "A Theory of Ice Ages, II," *Science,* 127, 3307 (1958), 1159–62.

FRITTS, HAROLD C., "Growth-Rings of Trees: Their Correlation with Climate," *Science,* 154, 3752 (1966), 973–79.

LADURIE, EMMANUEL LEROY, *Times of Feast, Times of Famine,* trans. by Barbara Bray. Garden City, N.Y.: Doubleday & Company, Inc., 1971.

LAMB, H. H., *Climates of the Present, Past and Future, Vol. 1, Fundamentals and Climate Now.* London: Methuen & Co. Ltd., 1972.

MITCHELL, J. MURRAY, ed., *Causes of Climatic Change,* Met. Monographs, Vol. 8, No. 30. Boston: American Meteorological Society, 1968.

NAMAIS, JEROME, "Short-Period Climatic Fluctuations," *Science,* 147, 3659 (1965), 696–706.

NEWMAN, JAMES E., "Climatic Changes: Some Evidences and Implications," *Weatherwise,* 24, 2 (1971), 56–62.

PETERSON, JAMES T., AND REID A. BRYSON, "Atmospheric Aerosols: Increased Concentrations during the Last Decade," *Science,* 162, 3849 (1968), 120–21.

PLASS, GILBERT N., "Carbon Dioxide and Climate," *Scient. Am.,* 201, 1 (1959), 41–47.

SCHWARZBACH, MARTIN, *Climates of the Past, An Introduction to Paleoclimatology,* trans. and ed. by Richard O. Muir. Princeton, N.J.: D. Van Nostrand Co., Inc., 1963.

SHAPLEY, HARLOW, ed., *Climatic Change.* Cambridge, Mass.: Harvard University Press, 1953.

chapter seventeen

Modification of Weather and Climate

In view of the significance of weather and climate in human activities it is logical to inquire whether a reasonable measure of control over atmospheric processes might be achieved, in effect producing "weather to order." The previous chapter reviewed some of the evidence that climates have been changed by natural causes. The possibility of duplicating those causes suggest modification on a grand scale. Accounts of attempts at rainmaking persist throughout the history of man, from ritual rain dances, through unmitigated hoaxes and trial-and-error experimentation, to modern theoretical approaches. Many people continue to equate weather modification with rainmaking, but its scope is much broader, encompassing any change in atmospheric processes wrought by man; because climate is an aggregate of weather such changes imply climatic effects to some degree. In this final chapter we shall examine some of the theoretical bases of atmospheric modification, experimental approaches, and attendant problems. It will be obvious that these topics entail a review of the physical principles that govern weather and climate.

MODIFYING MICROCLIMATES

A fundamental consideration in weather modification is that of scale. Changes on a small scale are relatively easy to initiate and control, yet their cumulative effects locally and their possible extension to areas of larger scale are difficult to assess. Man has been modifying his microclimatic environment ever since he first sought the shade of a tree, built a fire, or fashioned a shelter. He has the capacity to create microclimates for the benefit of himself, his animals, and cultivated plants, but he also destroys microclimates, often to the detriment of plant and animal life and sometimes to the distress of his neighbors. Removal of vegetation changes the albedo of the land surface, thereby affecting heat and moisture exchanges. Cutting of forests reduces the amount of rain or snow intercepted above ground level and alters the processes of evapotranspiration and runoff as well as the flow of air. Irrigation, frost prevention techniques, and the creation of windbreaks are representative agricultural practices that influence microclimates. (See Chapter 12.)

Buildings that are designed for control of their internal climates also affect the microclimates in their vicinities by influencing air movement, heat exchange, and moisture flux. Anyone who has contrasted the microenvironment over a hot pavement with that over a green lawn has a practical appreciation of man's influence. Roads are commonly drier than adjacent land, have a lower albedo, and are likely to be travelled by sources of atmospheric pollutants. The construction of an embankment across a small valley can disrupt air flow and create a distribution of temperature that is reflected in the composition of plant communities. The effects of factories, parks, hedges, swamp drainage, and umbrellas are additional random examples that illustrate the ways in which man and his works modify climate on a small scale. Any change in albedo, water capacity and retention, evaporation, transpiration, or surface roughness may produce a change in climate, but the results are complex, difficult to measure, and not easily predicted.

CITY CLIMATES

Cities concentrate people and their activities in small areas, thereby providing excellent opportunities to examine cultural modifications of climate. Although atmospheric pollution, which normally attains greater densities over cities, has received a large share of popular attention in recent years, it is not the only factor influencing city climates. Urban areas also differ from their rural counterparts in surface materials, surface shapes, and heat and moisture sources. In turn these affect radiation, visibility, temperature, wind, humidity, cloudiness, and precipitation. Table 17.1 summarizes the influence of urban environments on a number of climatic elements.

Concentrations of pollutants in the air above a city create an *urban aerosol,* which attenuates insolation, especially when the sun angle is

TABLE 17.1
Approximate Average Effects of Urbanization on Climatic Elements*

Elements	*Comparison with rural environment*
Pollutants	
Solid particles	10 times more
Gases	5 to 25 times more
Cloud cover	5 to 10 percent greater
Fog, winter	100 percent more
Fog, summer	30 percent more
Precipitation	5 to 10 percent more
Rain days with less than 5 mm	10 percent more
Relative humidity, winter	2 percent less
Relative humidity, summer	8 percent less
Radiation	15 to 30 percent less
Ultraviolet radiation, winter	30 percent less
Ultraviolet radiation, summer	5 percent less
Duration of sunshine	5 to 15 percent less
Annual mean temperature	0.5° to 1.0°C higher
Heating degree days	10 percent fewer
Annual mean windspeed	20 to 30 percent less
Calms	5 to 20 percent more

*After Landsberg, Chandler, Lowry, and others.

low as is the case at high latitudes and in winter. The aerosol is best developed (that is, at its worst) during conditions of stable air and calms or light winds. (See section on Atmospheric Pollution in Chapter 13.) In comparison with open rural areas, the annual total direct solar radiation in the heart of large industrial cities may be decreased by 15 to 30 percent. Insolation has been observed to vary during the week, being greatest on Sundays, when industrial activity and traffic are at a minimum. The urban aerosol is somewhat selective, for it reduces the proportion of ultraviolet radiation more than the longer wave lengths. It reduces the number of bright sunshine hours as well as the horizontal visibility.

In spite of the diminished insolation the center of the typical

TABLE 17.2
Annual Mean Excess of Urban over Rural Temperatures for Selected Cities, C°*

Berlin	1.0	New York	1.1
Chicago	0.6	Paris	0.7
London	1.3	Philadelphia	0.8
Los Angeles	0.7	Washington	1.0
Moscow	0.7		

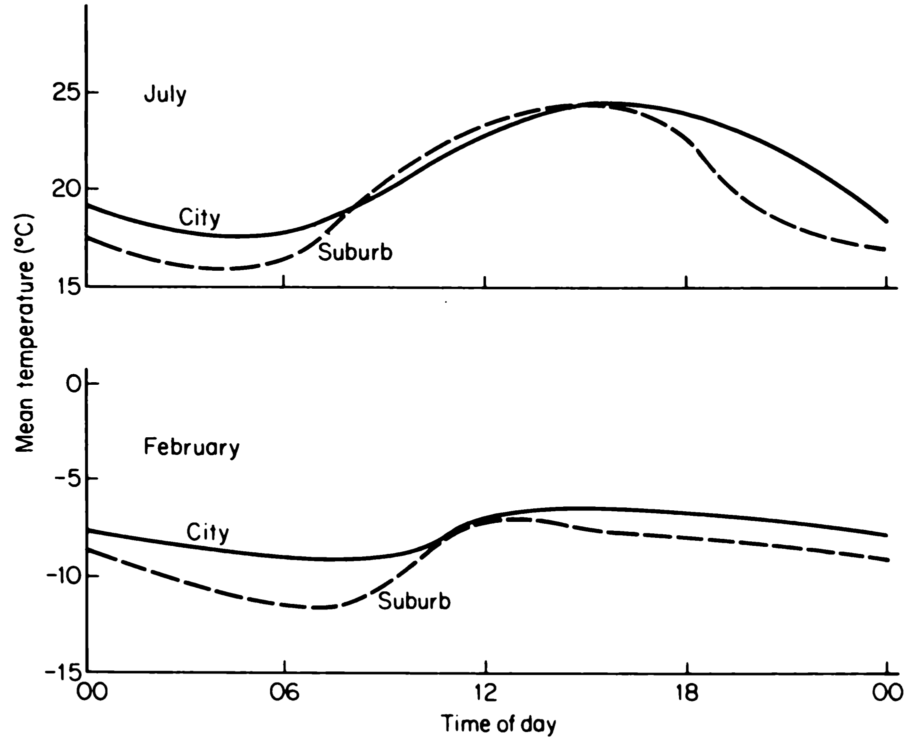
*James T. Peterson, *The Climate of Cities: A Survey of Recent Literature* (Raleigh, N.C., U.S. Dept. of Health, Education, and Welfare, 1969), p. 11.

metropolis constitutes a "heat island" that has a shape and size related to urban morphology, buildings, and industries and that results largely from urban heat generation and storage. Temperatures are normally highest near the city center and decline gradually toward the suburbs, beyond which there is a steep downward temperature gradient at the rural margin. The differences are greater at night than by day. Table 17.2 shows the annual mean positive temperature anomaly for several large cities. Although heat islands tend to be larger and more intense over large urban areas, there is not a direct relationship. Spacing of buildings and both kind and amount of activity also influence heat island development.

Owing to the blanketing effect of pollutants on the radiation budget, diurnal ranges of temperature are less over urban areas than over the countryside. Figure 17.1 illustrates typical daily temperature curves for a mid-latitude city and suburb. In view of the importance of vertical temperature lapse rates to atmospheric stability it is significant that nighttime inversions tend to be weaker over cities, where the heat island generates modest convection. To the lee of cities an urban heat plume at several meters above the surface may act to intensify rural inversions.

The roughness of the city surface increases frictional drag and

Figure 17.1
Diurnal temperature variations in a mid-latitude city and suburb. Data are for Schottenstift in urban Vienna (solid curve) and suburban Höhe Warte (dashed curve), 1956. (After Mitchell)



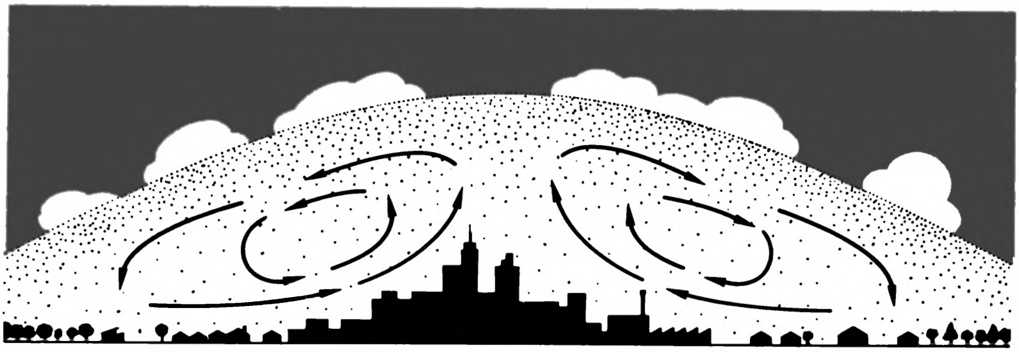


Figure 17.2
Schematic cross section of circulation generated by the heat island in an urban aerosol. (After Lowry)

turbulence. Gustiness and erratic flow of wind through the maze of urban canyons are well known to the city dweller, although gusts are more likely to reach their maximum speeds in the open countryside. Except under conditions of low regional wind speeds the mean wind speed within the city is lower than in the surrounding rural environment. When nighttime winds are light the speeds in the central city tend to be higher than in the country. Under nocturnal inversions the stable rural air inhibits surface flow, and calms are more frequent, whereas the relative instability of city air promotes turbulence, and stronger winds from above reach the surface more often. (Compare with the section on Diurnal Variation of Wind Speed in Chapter 4.)

A strong heat island generates its own circulation system (Figure 17.2). The inflow of cooler rural air toward the rising air over the city is generally weaker than might be expected, however, and is best developed on relatively calm, clear nights. Smoke plumes of residential areas have been observed to point toward a city center. Instrumental observations from towers and balloons have confirmed upward movement of air above urban heat islands.

The tendency of air to rise above the heat island is a possible explanation of greater cloudiness over cities and may account in part for greater precipitation, for cities are good sources of condensation and ice nuclei. One of the problems in urban precipitation studies is the difficulty of finding unpolluted rural areas with which to make experimental comparisons. A small number of ice nuclei injected into supercooled clouds might enhance rainfall, but a massive addition of condensation nuclei to air before it reaches its saturation temperature and produces warm clouds should lead to formation of many small droplets and inhibit rainfall. The net effect of the urban aerosol on precipitation over the city and to the leeward is not clearly understood. It depends on the kind and number of nuclei emitted by urban sources, natural nuclei, relative humidity, thermal lapse rates, and mechanical turbulence. Although climatological evidence is sparse, it may be that the incidence of hail is relatively greater over cities owing to convective activity. The proportion of precipitation in the form of snow appears to be less over urban centers, presumably because of higher temperatures associated with the heat island, and when it falls its effect on the surface albedo is rapidly modified by snow removal operations and soot. The effects of the heat island on turbulence and temperature suggest processes leading to increased total precipitation, but conclusive studies of a large number of cities under different conditions are lacking.

The mean relative humidity in city air is usually a few percent lower than in the surrounding country, especially at night and in summer when the heat island is well developed. Rural-urban differences in specific humidity are much more complex. Urban surfaces promote rapid runoff of precipitation, whereas vegetation and soil in the country retain moisture for evaporation over a longer period. On the other hand, the difference in availability of water is offset to some extent by the many combustion sources in the city. During periods of light winds tall buildings may inhibit the flow of air at ground level and thus reduce the upward diffusion of moist air. The net effect of all factors in a given city is to produce specific humidity values that reflect the form and function of the city.

INADVERTENT MODIFICATION OF MACROCLIMATES

The suggestion that human activities may unintentionally change world climates carries awesome implications that are frequently exaggerated beyond the limit of present evidence. Most of the linkages are theoretical; some are purely speculative. A few persistent ones deserve further examination.

The role of carbon dioxide as a possible factor in climatic changes was discussed in Chapter 16. The burning of large quantities of fossil fuels during the last two centuries has delivered carbon dioxide and water vapor to the atmosphere. Estimates indicate an increase of carbon dioxide amounting to 10 to 15 percent in the past century, but the rate of increase is rising. The ability of carbon dioxide to intercept outgoing long-wave radiation is a logical explanation of increasing temperatures. Gilbert Plass and others advanced the carbon dioxide theory to explain slight global temperature increases during the period between 1885 and 1940. Since 1940, however, there has been a cooling trend which suggests

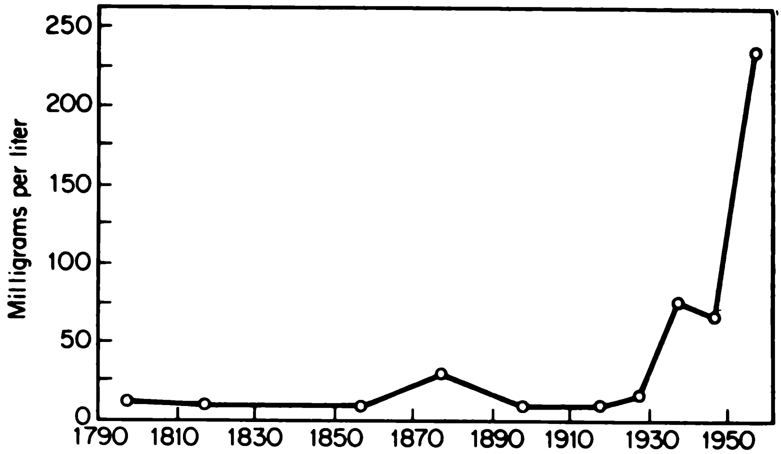


Figure 17.3
Amounts of dust measured in ice layers on the Caucasus Mountains, U.S.S.R. The small circles represent 10-year running means calculated from available samples. Explanations of the variations in dust accumulation include both local and global disturbances. (Adapted from F. F. Davitaya, "Atmospheric Dust Content as a Factor Affecting Glaciation and Climatic Change," *Ann. Assoc. Am. Geogrs.,* 59, 3 (1969), 552–60)

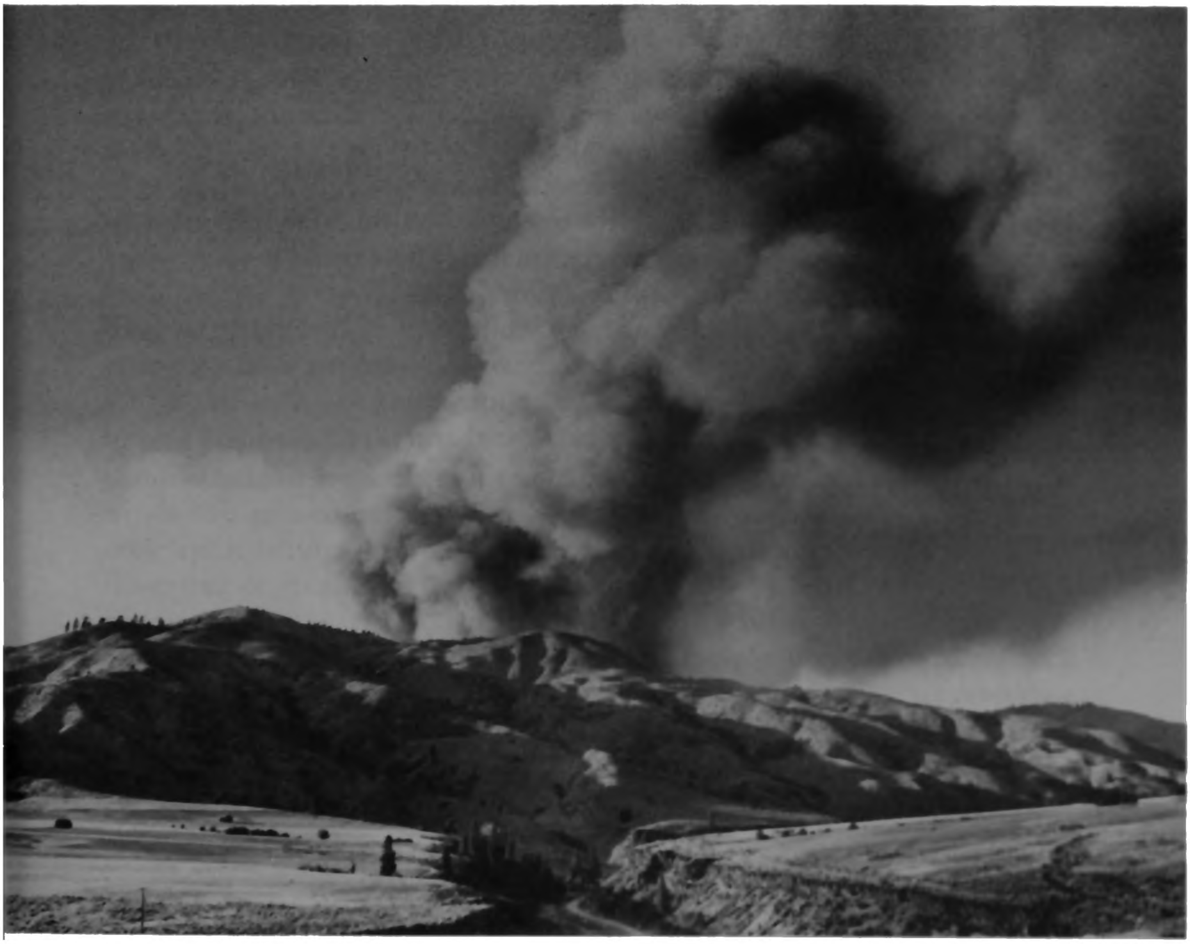
the influence of other factors. Much remains to be learned about the role of oceans, vegetation, and other factors in the carbon dioxide balance of the earth-atmosphere system.

Atmospheric dust is another possible link between man and climatic change. Basic theories concerning the effects of dust aerosols have been developed largely from observations following volcanic eruptions. Solid particles produced by man's activities do not normally reach to such heights, except in the cases of certain nuclear explosions and emissions from high-altitude rockets and aircraft. Within the troposphere the rate of return by gravity and precipitation is relatively rapid, and where the materials fall on ice or snow they may reduce the albedo and speed melting. Visual evidence and measurements indicate a worldwide increase in atmospheric dust, which reduces direct insolation and increases radiation from the sky. (See Figure 17.3.) The net effect of dust layers in the upper atmosphere is likely to be an increase in the planetary albedo and a consequent cooling of the lower atmosphere. Tropospheric dust of the types introduced by industrial and land-use practices, on the other hand, is more likely to absorb heat and lead to warming. (See Figure 17.4.) The effect of man-made dust is probably far less than that of volcanic eruptions.

Among the various kinds of solids emitted from surface sources into the atmosphere some appear to be more effective than others as nuclei for condensation or ice crystal formation. Lead particles in automobile exhausts, for example, become efficient ice nuclei when they combine with iodine or bromine. Too many hygroscopic particles, such as sea salt, tend to produce a large number of small cloud droplets that do not coalesce easily. The result could be a decrease in precipitation.

Emissions by high-flying aircraft and rockets of carbon dioxide, water vapor, and solid particles that act as freezing or condensation nuclei are possible causes of unintentional atmospheric modification. The amounts of these contaminants are relatively insignificant when viewed in terms of the total atmospheric volume and compared with more abundant surface emissions, but if they accumulate in the stratosphere they could alter the radiation balance. The effect of water vapor raises unique questions. It may be that condensation trails cause cooling by increasing the planetary albedo, although sky coverage of these artificial clouds is small and ephemeral, and high-level clouds should reduce radiative loss to space. It has also been suggested that cirriform jet trails precipitate ice crystals that could modify lower clouds, but the

Figure 17.4
Forest fires are among the many sources of atmospheric dust. (U.S. Forest Service)



confirming evidence is sparse. Even less is known about the effects of stratospheric flight on the ozone layer and about the influence of sonic waves on cloud dynamics. A decrease in ozone, which might result from interaction with nitric oxide emissions, would allow greater amounts of ultraviolet radiation to reach the earth's surface.

Exhausts from large rocket engines can introduce water vapor well into the thermosphere in quantities sufficient to disturb the radiation balance of the upper atmosphere. Nuclear explosions have injected quantities of radioactive debris into the stratosphere. The special significance of contaminants in the stratosphere and above is that they remain for periods of months or years before being lost to space or returned to earth, whereas in the troposphere most particles can be expected to settle or be washed out by precipitation within a few days.

Whether redistribution of surface waters has widespread climatic effects is a question with broad implications. It is known that large reservoirs have influences on the adjacent climate, reducing temperatures, increasing humidity, and sometimes generating local breezes. What is less certain is the extent to which regional precipitation is affected. Present knowledge does not lend strong support to claims that impounded water and irrigation projects lead to substantial increases in rainfall. Irrigated vegetation has a lower albedo than typical arid land surfaces, and although local temperatures are reduced by evaporation of irrigation water the net global effect may be a temperature increase owing to decreased reflection of insolation.

A form of pollution which would appear to have direct effects on climate is excess heat. The heat released by various kinds of fuel combustion, consumption of electrical power, mechanical friction, and even human bodies clearly influences microclimates. This raises the question of whether sustained thermal pollution from many sources might affect regional climates. Although the annual production of such energy equals only about one twenty-fifth of one percent of the total net solar radiation at the earth's surface, it is increasing at an increasing rate that warns of a future problem, especially in highly urbanized areas.

THEORETICAL BASES OF PLANNED MODIFICATION

The history of weather modification was dominated for many years by unsystematic experiments that were difficult to evaluate and often potentially dangerous. In recent decades the emphasis has shifted to attempts at controlled experimentation based on theoretical knowledge of atmospheric processes. The hypothetical possibilities include alteration of the heat and moisture budgets and the motion systems of the atmosphere and oceans. An attack on one necessarily has effects on the others. We shall defer discussion of the wisdom of attempting modification until the end of the chapter.

To examine theoretical approaches to modification of the atmosphere's energy budget we need review only the basic principles treated in Chapter 2 and the theories of natural climatic change in Chapter 16. At the outset we can consider the sun as the source of earth heat and reject for the foreseeable future any thoughts of changing the solar constant by altering such factors as the solar fire, the shape of the earth's orbit, inclination of the polar axis, or the rate of earth rotation. Partial eclipses of the sun or diversion of insolation to desired latitudes or to the night hemisphere might be achieved by means of satellites or clouds of reflective particles, but the procedures would strain present technology.

The potential effects of changing the types and amounts of dust and gases in the atmosphere have been discussed in the previous section. Materials that affect the atmospheric albedo and transmissivity of radiation at different wave lengths would undoubtedly alter the energy exchange. Rockets and satellites could inject water, smoke, metallic particles, carbon dioxide, and other substances into the atmosphere at reasonably controlled heights, but, as we have seen, the effects are uncertain.

More is known about the processes of heat exchange at and near the earth's surface, and the prospects are brighter for modification on limited scales with less risk of disaster. Serious proposals to alter the surface albedo range from dusting snow and ice with carbon black to enhance melting to the application of white powders for the purpose of reducing temperature and evaporation. It has been suggested that global changes in climate might be initiated by dusting the arctic ice pack with dark powder for a number of years. An ice-free Arctic Ocean would produce a vastly different regional exchange of heat and moisture. Proponents of the scheme claim that it would ameliorate the polar climates of the Northern Hemisphere. On a smaller scale the darkening of surfaces in arid areas would increase thermal convection and, according to theory, stimulate cloud formation and precipitation. Applied in the path of moist winds this procedure would create a "thermal mountain" and simulate orographic lifting. The most obvious approach to heat budget regulation is heating or cooling at the surface. Although heat losses from buildings, factories, compost heaps, orchard heaters, and other sources cause local temperature increases, the amount of energy required to effect regional or global heating by direct means far exceeds technological capabilities. The problems of regional cooling are at least as great. Only if a local triggering action were to induce processes on a larger scale could we expect such methods to be fruitful.

Precipitation, evapotranspiration, and runoff are the phases of the hydrologic cycle that offer possibilities for "control" of the moisture budget. An incomplete understanding of the processes of condensation, sublimation, coalescence of water droplets, and growth of ice crystals is the rather shaky foundation for modification of clouds, precipitation, and fog. Experimental tests and attendant theories will be treated separately in the following section. Meanwhile, consider the process of evaporation. One way to increase evaporation is to heat water. Among the suggestions has been the use of atomic power to heat known sources of atmospheric moisture; this presumably would lead to greater precipitation to the leeward. Another approach is the creation of large water bodies by means of dams or engineered stream diversions. Whereas artificial lakes clearly increase local evaporation, it is not equally evident that there is a consequent increase in precipitation either locally or to the leeward on a regional scale. Nor is the effect of evapotranspiration from irrigated lands and vegetated areas upon precipitation adequately understood. Proposals to plant deep-rooting phreatophytes in semiarid areas in order to bring ground water into the atmospheric phase of the water cycle have not been tested sufficiently, but there is scant evidence of their potential success in modifying climate.

The origin of water that is subsequently precipitated over land is still an unresolved question. Considering the restless nature of the troposphere and the vigor of winds, a molecule of water that leaves the surface in one area is not likely to be precipitated again until it has been transported hundreds or thousands of kilometers. Even conceding that humid climates tend to perpetuate themselves in part by the regional recycling of water, the enhancement of rainfall in arid climates by introducing reservoirs, irrigation, and vegetation has doubtful potential. The partial return of evaporated surface water as precipitation in a humid climate probably occurs for the simple reason that atmospheric conditions favor precipitation there. Arid climates are dry because the requisite dynamic processes are lacking. The example of the Red Sea reveals no direct cause and effect relationship between large water bodies and precipitation. Yet there have been numerous plans to transform the climate of the Sahara by creating a huge lake in its basin areas. The reservoir behind Aswan Dam in Egypt has not produced appreciable increments of rainfall in the downwind region of the central Sahara. These and other examples of deserts adjacent to water expanses leave little doubt that air mass subsidence in the general circulation is a more potent factor than evaporation. In regions of greater atmospheric instability extensive evaporating surfaces afford better possibilities for increasing precipitation by economically significant amounts.

Where water is scarce it may be more profitable to conserve it by reducing rather than promoting evapotranspiration. On a small scale this can be achieved by weed control and the use of mulches and by strategically placing windbreaks so as to reduce air flow across moisture sources. A technique of increasing importance on small reservoirs and lakes is the use of substances which form monomolecular layers on the surface and inhibit the escape of water molecules. Nontoxic alkanols such as hexadecanol and octadecanol have been employed for this purpose. The objections to oil or other polluting substances are obvious. In addition to retarding evaporation the surface film, or monolayer, suppresses tiny waves or ripples generated by light wind shear and thereby reduces upward diffusion of water vapor. Under conditions of moderate to high wind speeds, however, breaks in the monolayer render it ineffective.

Runoff control by means of engineering projects and the management of forests, grasslands, and agriculture needs no elaboration here, but the control of snowmelt deserves special mention. The application of carbon black or other dark materials to speed melting of ice and snow has already been suggested in connection with modification of the energy balance. A method that would retard melting could yield economic benefits by extending the availability of water into dry periods. Although no feasible way is known to reduce melting on a large scale, limited control might result from the spreading of reflective substances on dirty snowfields.

The simplest approaches to direct interference with air movement employ barriers and mechanical forcing. A redistribution of mountains would surely affect climate. Windbreaks and fans accomplish modest results in various agricultural practices. Increasing friction over large areas would decrease the surface winds of the same areas, although turbulence might be increased to an unacceptable extent. Schemes such as those to plant extensive shelterbelts on the North American Great Plains in the late nineteenth century and again in the 1930s might reduce surface winds, but the number of trees required for a regional amelioration of wind is so great as to require an overall change in climate. Proponents of shelterbelts have contended that trees would in fact alter the moisture and energy balances as well as winds.

The close relationship between thermal energy and motion is seen in the use of orchard heaters to induce convective activity in the surface inversion layer, in the circulation of the urban heat island, or in the altered wind patterns around large reservoirs. Proposals to assault major storms or segments of the general circulation are much more ambitious and in most cases highly speculative. Application of direct counterforces to divert violent storms or major wind systems is futile. Table 17.3 shows the immense amounts of energy developed in several types of atmospheric motion systems. Clearly it would not be feasible to concentrate the energy

TABLE 17.3
Estimated Kinetic Energy in Atmospheric Motion Systems and Time Required for all United States Generating Facilities to Produce Electrical Equivalent*

System	*Approximate Kinetic Energy (ergs)*	*Time Equivalent U.S. Electric Generation*
Tornado (life)	10^{21}	30 seconds
Small thunderstorm (life)	10^{22}	5 minutes
Large thunderstorm (life)	10^{23}	several hours
Hurricane (instantaneous)	10^{25}	several days
Extratropical cyclone	10^{26}	5 to 6 weeks
General circulation of Northern Hemisphere	5×10^{27}	6 years

******Weather and Climate Modification,* Report of the Special Commission on Weather Modification (Washington, D.C.: National Science Foundation, 1966), pp. 35–36.

required to overpower these phenomena. Large-scale manipulation of the energy exchange and initiation of triggering effects in conditions of natural instability offer greater promise.

A characteristic of the general circulation as we presently understand it is the irregularity of its perturbations. Consequently conditions that are quite similar initially may develop into very different physical states owing to slight nuances in physical processes. This suggests the possibility of triggering natural instabilities to effect changes in weather and, by continuing the triggering action, to alter climate. Because energy, moisture, and motion systems do not operate separately in the atmosphere, however, a relatively small disturbance of one may induce a chain reaction of complex processes that are difficult to predict and impossible to control.

Insofar as ocean currents and drifts influence climates, alteration of their speed or direction would initiate climatic changes. Mechanical mixing to produce artificial upwelling of cold water not only would affect heat and moisture exchanges at the sea surface but also would cause changes in the horizontal circulation. Thermal mixing could be achieved by using deep sea water as a coolant in nuclear and other thermal power generators and reintroducing it to seek higher levels in the ocean. The results of artificial mixing might be felt in the climates of areas far removed from the operation, however, and the atmospheric effects might be difficult to reverse.

One of the most persistent suggestions for climatic modification on a large scale is that to dam Bering Strait between Siberia and Alaska. If water were then pumped out of the Arctic Ocean into the North Pacific warmer water from the North Atlantic would be drawn into the arctic basin to melt the ice. This proposal, like that to melt the ice pack by reducing its albedo, is supposed to produce milder polar and tundra climates in the Northern Hemisphere. It is conceivable that it might raise temperatures, increase evaporation from the open ocean, promote greater cloudiness and snowfall, and trigger another ice age. It might also create stronger winter winds from the cold anticyclones over the continents, thus enhancing storminess. The relevant question appears to be not whether the climate would be altered but rather *how* it would be altered. Other theoretical possibilities for regulating oceanic circulation include damming the Florida Strait to modify the Gulf Stream and engineering the exchange of water through the Strait of Gibraltar. Accelerated breakup of the coastal ice around Antarctica would affect ocean currents at least in the Southern Hemisphere. Proposals to tow ice rafts to arid coasts to augment water supplies also entail heat transport. Removal of sufficient ice would increase the area of exposed water along the antarctic perimeter, probably creating climatic effects. Like many other suggestions for global engineering these ideas pose economic as well as scientific problems.

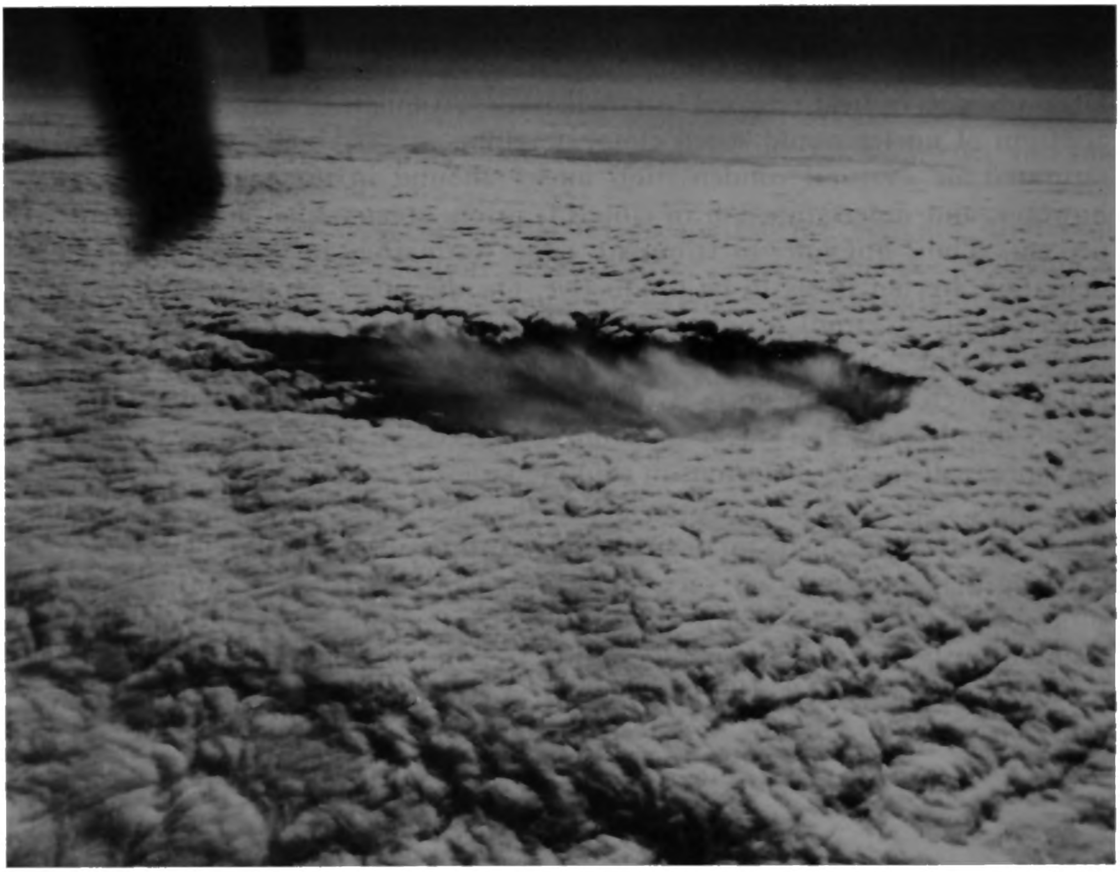
CLOUD MODIFICATION

The principle of a triggering action to initiate weather changes has had its most widespread applications in cloud modification experiments that seek to duplicate natural processes. The critical processes are condensation, sublimation, ice crystal formation, and coalescence. In view of the large amounts of heat released by condensation it might be infered that addition of nuclei would speed cloud development, specially in supersaturated air. Natural condensation nuclei abound in the atmosphere, however, and supersaturation in clouds is rarely greater than about one percent. Thus, atempts to stimulate cloud formation by forcing condensation are not likely to be productive. On the other hand, extensive supercooling of liquid droplets is common in natural clouds. Freezing releases latent heat of fusion and may disrupt the energy system of the cloud or even an entire storm. Growth of ice crystals and coalescence of liquid droplets are the stages immediately preceding precipitation and are more realistic considerations in cloud modification.

Most experimental work on artificial stimulation of precipitation has involved clouds made up of supercooled droplets. In 1945 Bohdan M. Cwilong demonstrated that very small droplets of pure water could be supercoled to — 40°C. Below that critical temperature frezing occurs spontaneously, but higher temperatures freezing takes place only when ice nuclei are present. During laboratory experiments in 1946 Vincent J. Schafer found dry ice to be efective in cooling droplets to produce ice crystals. He later sprinkled 3 kg of dry ice into a cloud from an airplane at 4,300 m above western Massachusetts. Snow fell from the cloud, although it did not reach the ground. This pioneer experiment led to widespread seeding with dry ice in cloud modification. The results have not been uniformly predictable, but observers have reported changes in cloud structure and falls of rain or snow in numerous experiments. (See Figure 17.5.)

A related technique of cloud seeding is artificial nucleation, the purpose of which is to provide sufficient ice nuclei to trigger precipitation or, by "overseeding," to create so many ice crystals that few become large enough to fall from the cloud. In either case artificially induced ice crystals are presumed to grow at the expense of water droplets and fall as precipitation. (See Chapter 3, Precipitation: Causes, Forms, Processes, and Types.) Natural nuclei initiate the process in supercooled clouds at temperatures between −40° and −15°C; certain dust particles from the North American Great Plains are effective at temperatures as high as −7°C; crystals of silver iodide start ice crystal formation at temperatures between approximately −13° and −4°C. Ice appears to form more

Figure 17.5
Hole formed in supercooled clouds following dry ice seeding near Goose Bay, Labrador. (Air Force Cambridge Research Laboratories)



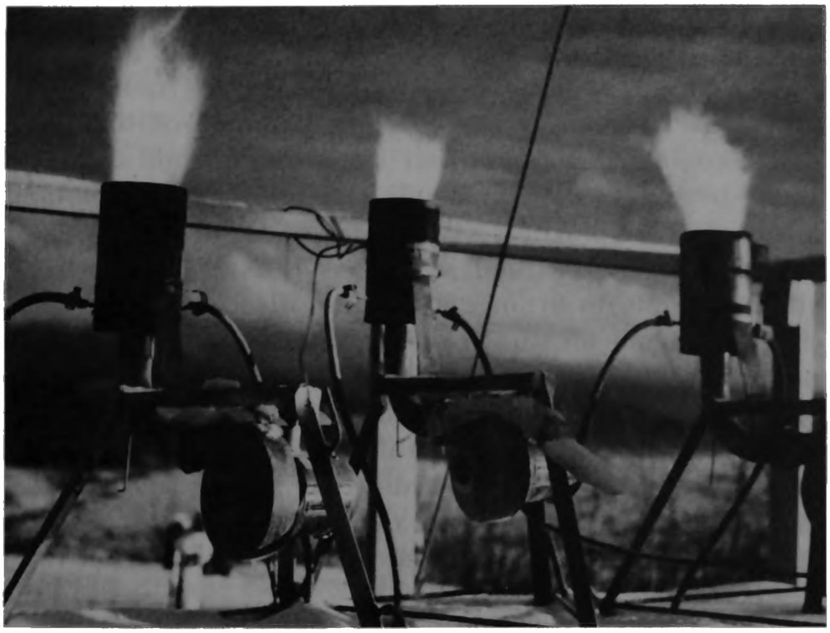


Figure 17.6
Silver iodide generators in operation. The burning mixture contains silver iodide, acetone, and propane. (Bureau of Reclamation, U.S. Department of the Interior)

readily upon silver iodide crystals because the crystalline structures of both are hexagonal, and they are of comparable size. Thus, silver iodide can be effective in warmer (though supercooled) clouds and at lower (warmer) levels in a cloud. Silver iodide has another advantage in that it can be heated and released at low levels as a vapor which quickly reverts to crystals that ascend in convection currents. Ground-based experiments to induce precipitation in specific regions make use of generators to dispense silver iodide from strategic upwind locations. (See Figure 17.6.) Stable air lying over the area tends to suppress the rising crystals, however, and aircraft or rockets may be needed to reach the target clouds.

Another application of silver iodide seeding aims to produce dynamic effects in supercooled cumulus. Massive seeding with 100 to 1,000 times as many nuclei as used in ordinary experiments results in rapid release of sufficient heat of fusion to promote vertical cloud development and enhance rainfall. Tests of the dynamic seeding hypothesis show great variation in results, but on the average seeded clouds yield more rain than control (unseeded) clouds.

Clouds having temperatures above 0°C throughout present special problems in cloud seeding. Low stratus, for example, is thermodynamically stable, and ice nucleation and release of latent heat of fusion are less significant processes than coalescence of droplets. Where droplets of varying sizes exist together their different rates of movement may lead to collision and coalescence. Large droplets may divide upon collision, creating several drops which set off a chain reaction of raindrop growth. Freezing of drops by means of dry ice might facilitate collision. Introduction of hygroscopic materials or water spray into a cloud could provide the necessary large drops. Warm clouds have received less attention than supercooled types, and practical means of modification are lacking.

The effectiveness of cloud seeding is difficult to appraise. It is practically impossible to arrange a scientifically controlled experiment in which two or more identical clouds are available for different treatment and observation of results. The alternative is random selection from among a number of similar clouds and statistical comparisons of many experiments with each other and with climatic probabilities. Claims of changes in precipitation by cloud seeding vary widely and include both increases and decreases. Methods of evaluation also differ, but proponents of precipitation enhancement by seeding report increases on the order of 10 to 20 percent.

In summary, the possible results of seeding convective clouds are many and varied. A 1966 report of a special National Academy of Sciences Panel on Modification of Weather and Climate listed the following possible induced responses to seeding:

1. Increase or decrease of rain, snow, or hail
2. Increase or decrease of precipitation leeward of the seeded area
3. Increase or decrease of lightning
4. Increased buoyancy and vertical cloud development
5. Altered patterns of air movement that may affect adjacent clouds

Thus, we face two fundamental questions: What kind and amount of human intervention will set changes in motion, and what are all the possible results of such action?

In addition to cloud seeding, electricity and shock waves offer theoretical possibilities in cloud modification. Lightning produces heat which influences convective activity, and electrical charges are probable factors in the coalescence of tiny cloud droplets. These effects are plausible explanations for the gushes of precipitation which often follow a lightning discharge in cumulonimbus. Alteration of the electrical field by artificial means could affect precipitation. Thunder creates shock waves that may induce collision of droplets, and the associated cooling by expansion may create ice crystals that assist the development of precipitation. Beginning in the nineteenth century cannonading was employed in rainmaking attempts. In more recent times experimental rocket explosions within clouds have yielded few conclusive results.

MODIFICATION OF STORMS

As the scale of atmospheric modification increases the complexity of physical processes and the uncertainties of the outcome also increase. We have seen that counterforces are impractical solutions for the abatement of major storm systems. Most experiments in the modification of storms are directed toward changing their dynamic characteristics, mainly by cloud seeding. If, as some meteorologists believe, thunderstorms and tornado vortices draw critical portions of their energy from the heat of electrical discharges, then seeding at early stages of development could disrupt the energy balance and forestall violent convection. It has even been suggested that small strips of metal dropped into thunderstorms having high concentrations of electrical charges might decrease the potential for tornado development. The great number of lightning-caused forest fires in the western United States has prompted a major U.S. Forest Service research project in lightning suppression. Known as Project Skyfire, its aim has been the modification of the electrical field in thunderstorms by silver iodide seeding on the supposition that an abundance of ice crystals increases the discharge points in the cloud and suppresses ground strikes. The use of metallic chaff is based upon a similar hypothesis. In other experiments small rockets have been sent aloft to carry long, trailing wires to induce lightning.

Hail suppression experiments are closely related to thunderstorm modification. One approach is seeding to create more small ice particles and more but smaller hail stones. Another is to seed the supercooled layers copiously, thus reducing the number of supercooled droplets available for ice accretion. Still other techniques rely on the theory that hail formation begins on large water drops near the base of the cloud, where added quantities of condensation nuclei might produce smaller drops. Russian scientists have claimed remarkable accuracy in radar detection of cloud layers in which hailstones are being formed and into which silver iodide can be seeded by artillery shells.

Cloud seeding has also been the fundamental method in tropical cyclone modification experiments. The objective has been primarily to reduce wind speeds in the storm center on the theory that heat of fusion added as a result of the seeding would alter the pressure field and lead to enlargement of the eye. Observed wind speeds decreased following seeding experiments in Atlantic hurricanes in 1961, 1963, and 1969 under Project Stormfury, a joint effort of the United States Navy and the National Oceanic and Atmospheric Administration. Whether seeding or natural changes in the hurricanes led to the decrease is not certain. Nor is there sufficient evidence to indicate that storms could be steered by seeding. After an early seeding attempt under Project Cirrus in 1947 a hurricane suddenly changed course and moved across Savannah, Georgia, causing heavy damage. The 1971 seeding of Hurricane *Ginger* was followed by damage in North Carolina. Scientists who would modify hurricanes are understandably cautious.

Inasmuch as hurricanes derive their energy primarily from the sea surface it is logical to assume that reduction of evaporation might prevent development, or at least calm their fury. Alkanols, which may be relatively effective on the nearly calm waters of lakes and reservoirs, offer little promise for suppression of evaporation from the disturbed waters beneath a developing hurricane. Oil could be expected to form a tougher film, but it would also reduce the albedo slightly, thus partially offsetting its capacity to retard evaporation. The polluting characteristics of oil slicks might be alleviated by using biodegradable synthetic oils.

It is tempting to consider jet streams, fronts, and pressure systems as direct targets for modification. However, current knowledge of the physical processes that guide these phenomena and their irregularities is inadequate for construction of reliable models for their transformation.

FOG DISPERSAL

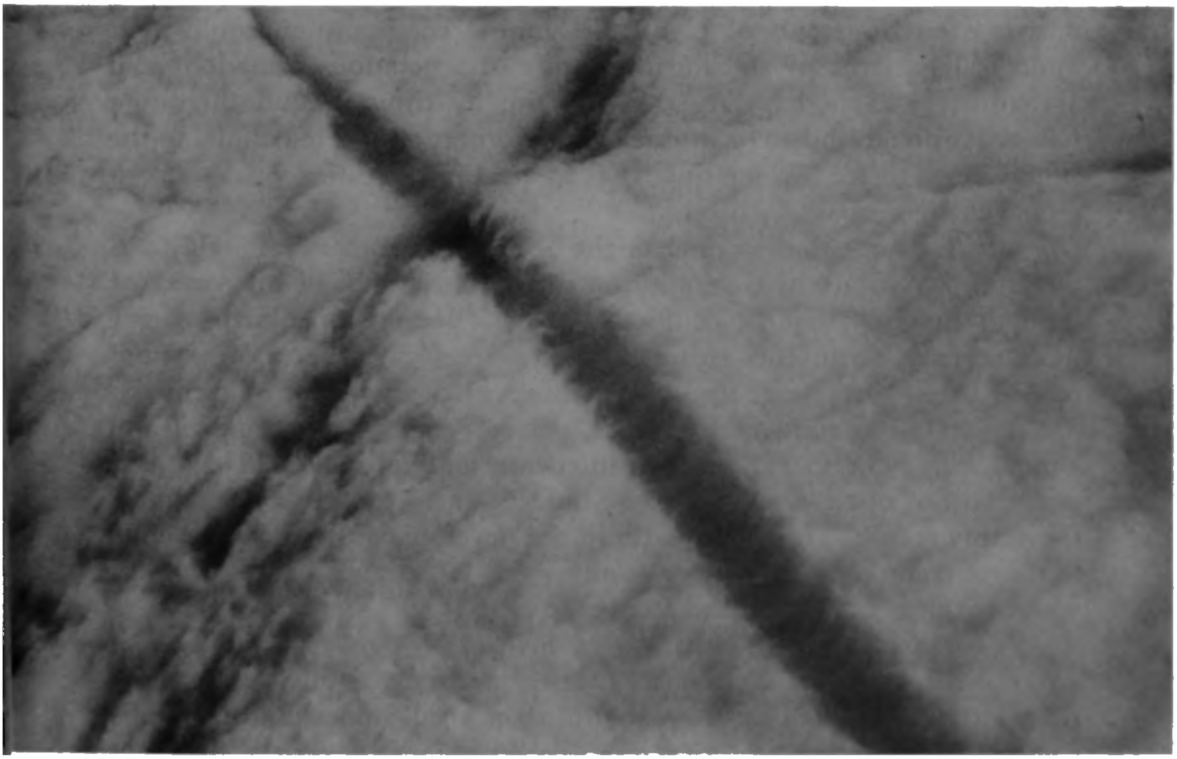
Because fog lies at the earth's surface both the techniques and evaluation of modification are somewhat simpler than for clouds, but fog, like clouds, occurs in both supercooled and warm forms. The most successful attempts to modify cold fogs have employed seeding procedures similar to those applied to supercooled clouds. Dry ice seeding by small aircraft, flying from fog-free air strips, has been highly if not uniformly effective in clearing fog at a number of major airports. Another cooling method uses liquid propane, which is sprayed into the fog from ground dispensers. As the proprane vaporizes the gas expands and cools adiabatically, thus reducing the temperature of the supercooled droplets below the critical minimum for the liquid state.

Warm fogs present many of the same problems encountered in the modification of warm clouds. They are not only more difficult to dissipate than cold fogs, but they are also more frequent and widespread. Warm advection fogs are continually renewed over a given area such as an airport, whereas stable ground fogs tend to form in conditions of near calm. Early successes in ground fog dispersal were achieved by heating. During World War II the British developed FIDO (Fog Investigation and Dispersal Operation), which evaporated fog by burning oil dispensed from perforated pipes along runways. Military expediency justified the high cost, but the principle of fog dispersal by evaporation was physically sound. Heating and turbulence created by jet engine exhaust, simple combustion blowers, or other sources of energy have potential for removal of shallow warm fogs. (See Figure 17.7.) Seeding with carbon black is intended to promote evaporation by heat gained from the absorption of radiant energy. A less direct approach to forced evaporation is the use of highly hygroscopic substances which remove water vapor from the air within the fog and thereby cause fog droplets to evaporate. Calcium chloride, in powdered form and as a spray solution, was used in early experiments, but its corrosive properties are objectionable.

Japanese scientists were among the first to test water seeding of warm fogs; occasionally they achieved temporary clearing as the result of drop collision and precipitation. Other attempts to induce coalescence employ electrical charges introduced on oil spray or by means of wires. So-called "fog brooms" consist of filaments of nylon or noncorrosive wires on frames placed so as to intercept droplets. Experiments based on this principle have been designed both to improve visibility in advection fogs and to capture water along desert coasts.

The cost of fog to commercial aviation and other forms of transportation is adequate justification for experimental investigation of fog

Figure 17.7
Fog dissipation in the wake of a Boeing 747 aircraft landing at Seattle-Tacoma airport. Entrainment of dry air from above enhances dissipation by heat of the jet engines. (Courtesy Stinson Aircraft, Inc.)



dispersal. All known techniques are expensive and must be weighed against losses that could otherwise be expected.

IMPLICATIONS OF WEATHER AND CLIMATE MODIFICATION

Thus far we have confined ourselves largely to the theories and experimental techniques of weather modification, and have avoided the pressing question of whether we should tamper with the atmosphere. Ability to forecast natural weather is a prerequisite to altering weather predictably. So long as the effects of modification activities are uncertain with respect to both time and scale, wisdom dictates discretion. Even if forecasts of natural atmospheric events were possible and the total meteorological results of human intervention were known, there remains the serious question of effects on other elements of the natural environment and on human activities. Small changes in the heat and moisture balances could produce far-reaching results by disrupting the ecological relationships of soils, plants, or animal life, all of which depend on the atmosphere. For example, increased precipitation on a plant community might favor certain species over others, promote soil leaching and erosion, encourage new insects and diseases, and create problems of runoff control. Small but sustained temperature changes, especially in critically marginal environments of plant and animal communities, could alter biological patterns by shifting the limits of vegetation formations and agriculture and perhaps interrupting the distribution and migratory habits of animals, birds, or insects. The imagination sets few limits to the implications of ice cap melting, rises in sea level, creation of deserts, or global changes in wind patterns.

If we assume that weather might be modified under reasonably predictable conditions that permit control of duration and area affected we face the challenge of conflicting needs. Because one man's picnic weather may be another's drought there are legal aspects of weather modification. They range through questions of cloud ownership, property damage liability, accident insurance, private versus public control and benefit, the right to experiment, interstate and international conflicts, and a host of other social, economic, and political problems. Legal cases inevitably reached the courts early in the development of cloud seeding, but a coherent body of weather modification law emerged slowly to deal with planning, execution, and after-effects.

The prospect of modifying our environment by manipulation of atmospheric processes is at once alluring and frightening. Its implications are no less important to mankind than a full understanding of the complex atmosphere.

Additional Readings

BLACK, JAMES F., "Weather Control: Use of Asphalt Coatings to Tap Solar Energy," *Science,* 139, 3551 (1963), 226–27.

DAVITAYA, F. F., "Atmospheric Dust Content as a Factor Affecting Glaciation and Climatic Change," *Ann. Assoc. Am. Geogrs.,* 59, 3 (1969), 552–60.

FLEAGLE, ROBERT G., ed., *Weather Modification: Science and Public Policy.* Seattle: University of Washington Press, 1969.

HICKS, J. R., "Improving Visibility near Airports during Periods of Fog," *Journ. Appl. Met.,* 6, 1 (1967), 39–42.

Inadvertent Modification of Climate, Report of the Study of Man's Impact on Climate. Cambridge, Mass.: The MIT Press, 1971.

LANDSBERG, HELMUT E., "Man-Made Climatic Changes," *Science,* 170, 3964 (1970), 1265–74.

LOWRY, WILLIAM P., "The Climate of Cities," *Scient. Am.,* 217, 2 (1967), 15–23.

MALKUS, JOANNE S., and ROBERT H. SIMPSON, "Modification Experiments on Tropical Cumulus Clouds," *Science,* 145, 3632 (1964), 541–48.

McDONALD, JAMES E., "The Evaporation-Precipitation Fallacy," *Weather,* 17, 5 (1962), 168–77.

MITCHELL, J. MURRAY, JR., "The Temperature of Cities," *Weatherwise,* 14, 6 (1961), 224–29.

MORRIS, EDWARD A., "The Law and Weather Modification," *Bull. Am. Met. Soc.,* 46, 10 (1965), 618–22.

National Research Council, Committee on Atmospheric Sciences, *Weather and Climate Modification: Problems and Progress.* Washington, D.C.: National Academy of Sciences, 1973.

PETERSON, JAMES T., *The Climate of Cities: A Survey of Recent Literature.* Raleigh, N.C.: U.S. Department of Health, Education and Welfare, 1969.

SEWELL, W. R. DERRICK, ed., *Human Dimensions of Weather Modification.* Chicago: University of Chicago, Department of Geography, 1966.

TRIBUS, MYRON, "Physical View of Cloud Seeding," *Science,* 168, 3928 (1970), 201–11.

Urban Climates, Tech. Note No. 108. Geneva: WMO, 1970.

WOODLEY, WILLIAM L., "Rainfall Enhancement by Dynamic Cloud Modification," *Science,* 170, 3954 (1970), 127–32.

Appendices

appendix a

Conversion Equivalents

LENGTH

1 kilometer = 0.6214 statute mile = 0.5396 nautical mile

1 statute mile = 0.8684 nautical mile = 1.6093 kilometers

1 meter = 1.0936 yards = 3.2808 feet = 39.37 inches

1 foot = 0.3048 meter

1 centimeter = 0.3937 inch = 10^4 microns

1 inch = 25.4 millimeters

SPEED

1 knot = 1 nautical mile per hour = 1.1516 statute miles per hour
= 0.5148 meters per second = 1.854 kilometers per hour

1 mile per hour = 0.8684 knot = 0.447 meters per second
= 1.609 kilometers per hour

1 meter per second = 2.2369 miles per hour = 1.9424 knots
= 3.2808 feet per second

MASS

1 kilogram = 2.2046 pounds (avoirdupois)
1 pound = 453.59 grams = 0.4536 kilogram
1 ounce = 28.3495 grams
1 gram = 0.0353 ounce

ENERGY

1 gram-calorie = 4.186 × 10^7 ergs = 4.186 joules
= 3.97 × 10^{-3} BTU
1 joule = 10^7 ergs = 0.738 foot-pound
1 BTU (British Thermal Unit) = 252 gram-calories
= 1,055 joules
1 erg = 2.388 × 10^{-8} gram-calories = 1 dyne cm
1 langley = 1 calorie per cm^2

appendix b

Abridged Meteorological Tables

TABLE B.1

Fahrenheit Equivalents of Celsius Temperatures (The column on the left contains 10C° intervals; the single digits across the top are 1C° intervals. Thus, 10°C = 50°F; 11°C = 51.8°F; 12°C = 53.6°F; and so on.)

Celsius	*Fahrenheit*									
	0	1	2	3	4	5	6	7	8	9
–50	–58.0	–59.8	–61.6	–63.4	–65.2	–67.0	–68.8	–70.6	–72.4	–74.2
–40	–40.0	–41.8	–43.6	–45.4	–47.2	–49.0	–50.8	–52.6	–54.4	–56.2
–30	–22.0	–23.8	–25.6	–27.4	–29.2	–31.0	–32.8	–34.6	–36.4	–38.2
–20	– 4.0	– 5.8	– 7.6	– 9.4	–11.2	–13.0	–14.8	–16.6	–18.4	–20.2
–10	14.0	12.2	10.4	8.6	6.8	5.0	3.2	1.4	– 0.4	– 2.2
– 0	32.0	30.2	28.4	26.6	24.8	23.0	21.2	19.4	17.6	15.8
+ 0	32.0	33.8	35.6	37.4	39.2	41.0	42.8	44.6	46.4	48.2
10	50.0	51.8	53.6	55.4	57.2	59.0	60.8	62.6	64.4	66.2
20	68.0	69.8	71.6	73.4	75.2	77.0	78.8	80.6	82.4	84.2
30	86.0	87.8	89.6	91.4	93.2	95.0	96.8	98.6	100.4	102.2
40	104.0	105.8	107.6	109.4	111.2	113.0	114.8	116.6	118.4	120.2

TABLE B.2

Celsius Equivalents of Fahrenheit Temperatures (The column on the left contains 10F° intervals; the single digits across the top are 1F° intervals. Thus, 10° F = -12.2° C; 11° F = -11.7° C; 12° F = -11.1° C; and so on.)

Fahrenheit	*Celsius*									
	0	1	2	3	4	5	6	7	8	9
-100	-73.3	-73.9	-74.4	-75.0	-75.6	-76.1	-76.7	-77.2	-77.8	-78.3
- 90	-67.8	-68.3	-68.9	-69.4	-70.0	-70.6	-71.1	-71.7	-72.2	-72.8
- 80	-62.2	-62.8	-63.3	-63.9	-64.4	-65.0	-65.6	-66.1	-66.7	-67.2
- 70	-56.7	-57.2	-57.8	-58.3	-58.9	-59.4	-60.0	-60.6	-61.1	-61.7
- 60	-51.1	-51.7	-52.2	-52.8	-53.3	-53.9	-54.4	-55.0	-55.6	-56.1
- 50	-45.6	-46.1	-46.7	-47.2	-47.8	-48.3	-48.9	-49.4	-50.0	-50.6
- 40	-40.0	-40.6	-41.1	-41.7	-42.2	-42.8	-43.3	-43.9	-44.4	-45.0
- 30	-34.4	-35.0	-35.6	-36.1	-36.7	-37.2	-37.8	-38.3	-38.9	-39.4
- 20	-28.9	-29.4	-30.0	-30.6	-31.1	-31.7	-32.2	-32.8	-33.3	-33.9
- 10	-23.3	-23.9	-24.4	-25.0	-25.6	-26.1	-26.7	-27.2	-27.8	-28.3
- 0	-17.8	-18.3	-18.9	-19.4	-20.0	-20.6	-21.1	-21.7	-22.2	-22.8
+ 0	-17.8	-17.2	-16.7	-16.1	-15.6	-15.0	-14.4	-13.9	-13.3	-12.8
10	-12.2	-11.7	-11.1	-10.6	-10.0	- 9.4	- 8.9	- 8.3	- 7.8	- 7.2
20	- 6.7	- 6.1	- 5.6	- 5.0	- 4.4	- 3.9	- 3.3	- 2.8	- 2.2	- 1.7
30	- 1.1	- 0.6	0.0	0.6	1.1	1.7	2.2	2.8	3.3	3.9
40	4.4	5.0	5.6	6.1	6.7	7.2	7.8	8.3	8.9	9.4
50	10.0	10.6	11.1	11.7	12.2	12.8	13.3	13.9	14.4	15.0
60	15.6	16.1	16.7	17.2	17.8	18.3	18.9	19.4	20.0	20.6
70	21.1	21.7	22.2	22.8	23.3	23.9	24.4	25.0	25.6	26.1
80	26.7	27.2	27.8	28.3	28.9	29.4	30.0	30.6	31.1	31.7
90	32.2	32.8	33.3	33.9	34.4	35.0	35.6	36.1	36.7	37.2
100	37.8	38.3	38.9	39.4	40.0	40.6	41.1	41.7	42.2	42.8
110	43.3	43.9	44.4	45.0	45.6	46.1	46.7	47.2	47.8	48.3

TABLE B.3

Millibar Equivalents of Inches of Mercury

Inches	*Millibars*									
	.0	.1	.2	.3	.4	.5	.6	.7	.8	.9
15	508	511	515	518	521	524	528	532	535	538
16	542	545	549	552	555	559	562	565	569	572
17	576	579	582	586	589	593	596	599	603	606
18	610	613	616	620	623	626	630	633	637	640
19	643	647	650	654	657	660	664	667	670	674
20	677	681	684	687	691	694	698	701	704	708
21	711	714	718	721	725	728	731	735	738	742
22	745	748	752	755	759	762	765	769	772	775
23	779	782	786	789	792	796	799	803	806	809
24	813	816	819	823	826	830	833	836	840	843
25	847	850	853	857	860	863	867	870	874	877
26	880	884	887	891	894	897	901	904	908	911
27	914	918	921	924	928	931	935	938	941	945
28	948	952	955	958	962	965	968	972	975	979
29	982	985	989	992	996	999	1002	1006	1009	1012
30	1016	1019	1023	1026	1029	1033	1036	1040	1043	1046
31	1050	1053	1057	1060	1063	1067	1070	1073	1077	1080

TABLE B.4
Dew Point Temperature (1,000 mb)

Dry-bulb temp. (°C)	*Saturation vapor press. (mb)*	*Wet-bulb depression ($T_d - T_w$)*																					
		1	*2*	*3*	*4*	*5*	*6*	*7*	*8*	*9*	*10*	*11*	*12*	*13*	*14*	*15*	*16*	*17*	*18*	*19*	*20*	*21*	*22*
-20	1.2540	-33																					
-18	1.4877	-28																					
-16	1.7597	-24																					
-14	2.0755	-21	-36																				
-12	2.4409	-18	-28																				
-10	2.8627	-14	-22																				
-8	3.3484	-12	-18	-29																			
-6	3.9061	-10	-14	-22																			
-4	4.5451	-7	-11	-17	-29																		
-2	5.2753	-5	-8	-13	-20																		
0	6.1078	-3	-6	-9	-15	-24																	
2	7.0547	-1	-3	-6	-11	-17																	
4	8.1294	1	-1	-4	-7	-11	-19																
6	9.3465	4	1	-1	-4	-7	-13	-21															
8	10.722	6	3	1	-2	-5	-9	-14															
10	12.272	8	6	4	1	-2	-5	-9	-14	-28													
12	14.017	10	8	6	4	1	-2	-5	-9	-16													
14	15.977	12	11	9	6	4	1	-2	-5	-10	-17												
16	18.173	14	13	11	9	7	4	1	-1	-6	-10	-17											
18	20.630	16	15	13	11	9	7	4	2	-2	-5	-10	-19										
20	23.373	19	17	15	14	12	10	7	4	2	-2	-5	-10	-19									
22	26.430	21	19	17	16	14	12	10	8	5	3	-1	-5	-10	-19								
24	29.831	23	21	20	18	16	14	12	10	8	6	2	-1	-5	-10	-18							
26	33.608	25	23	22	20	18	17	15	13	11	9	6	3	0	-4	-9	-18						
28	37.796	27	25	24	22	21	19	17	16	14	11	9	7	4	1	-3	-9	-16					
30	42.430	29	27	26	24	23	21	19	18	16	14	12	10	8	5	1	-2	-8	-15				
32	47.551	31	29	28	27	25	24	22	21	19	17	15	13	11	8	5	2	-2	-7	-14			
34	53.200	33	31	30	29	27	26	24	23	21	20	18	16	14	12	9	6	3	-1	-5	-12	-29	
36	59.422	35	33	32	31	29	28	27	25	24	22	20	19	17	15	13	10	7	4	0	-4	-10	
38	66.264	37	35	34	33	32	30	29	28	26	25	23	21	19	17	15	13	11	8	5	1	-3	-9
40	73.777	39	37	36	35	34	32	31	30	28	27	25	24	22	20	18	16	14	12	9	6	2	-2

TABLE B.5

Relative Humidity in Percent (1,000 mb)

Dry-bulb temp. (°C)	*Wet-bulb depression ($T_d - T_w$)* 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22
-20	28																					
-18	40																					
-16	48	0																				
-14	55	11																				
-12	61	23																				
-10	66	33	0																			
-8	71	41	13																			
-6	73	48	20	0																		
-4	77	54	32	11																		
-2	79	58	37	20	1																	
0	81	63	45	28	11																	
2	83	67	51	36	20	6																
4	85	70	56	42	27	14																
6	86	72	59	46	35	22	10	0														
8	87	74	62	51	39	28	17	6														
10	88	76	65	54	43	33	24	13	4													
12	88	78	67	57	48	38	28	19	10	2												
14	89	79	69	60	50	41	33	25	16	8	1											
16	90	80	71	62	54	45	37	29	21	14	7	1										
18	91	81	72	64	56	48	40	33	26	19	12	6	0									
20	91	82	74	66	58	51	44	36	30	23	17	11	5	0								
22	92	83	75	68	60	53	46	40	33	27	21	15	10	4	0							
24	92	84	76	69	62	55	49	42	36	30	25	20	14	9	4	0						
26	92	85	77	70	64	57	51	45	39	34	28	23	18	13	9	5						
28	93	86	78	71	65	59	53	47	42	36	31	26	21	17	12	8	4					
30	93	86	79	72	66	61	55	49	44	39	34	29	25	20	16	12	8	4				
32	93	86	80	73	68	62	56	55	46	41	36	32	27	22	19	14	11	8	4			
34	93	86	81	74	69	63	58	52	48	43	38	34	30	26	22	18	14	11	8	5		
36	94	87	81	75	69	64	59	54	50	44	40	36	32	28	24	21	17	13	10	7	4	
38	94	87	82	76	70	66	60	55	51	46	42	38	34	30	26	23	20	16	13	10	7	5
40	94	89	82	76	71	67	61	57	52	48	44	40	36	33	29	25	22	19	16	13	10	7

TABLE B.6

Supplementary Climatic Data

T : Temperature in degrees Celsius
P : Precipitation in millimeters

Latitude to nearest whole degree
Station height in meters

Africa		J	F	M	A	M	J	J	A	S	O	N	D	Yr
Accra, Ghana	T	27	28	28	28	27	26	25	24	25	26	27	27	26
6°N; 65 m	P	16	37	73	82	145	193	49	16	40	80	38	18	787
Addis Ababa, Ethiopia	T	17	18	19	19	19	17	15	15	16	16	17	17	17
9°N; 2,408 m	P	24	25	67	93	53	105	239	266	174	43	3	18	1,110
Algiers, Algeria	T	10	11	13	15	18	22	24	25	23	19	15	12	17
37°N; 28 m	P	116	76	57	65	36	14	2	4	27	84	93	117	691
Beira, Mozambique	T	27	28	27	26	23	21	20	21	23	25	26	27	24
20°N; 8 m	P	265	225	244	105	58	42	37	30	27	29	133	134	1,429
Bulawayo, Rhodesia	T	21	21	20	19	16	13	14	16	19	22	22	21	19
20°S; 1,344 m	P	134	111	65	21	9	3	T	1	5	25	89	125	589
Cape Town, South Africa	T	20	20	19	16	14	13	12	12	14	15	18	19	16
34°S; 44 m	P	11	15	14	53	89	84	83	73	45	31	17	11	526
Douala, Cameroon	T	27	27	27	27	27	26	25	25	25	26	26	27	26
4°N; 13 m	P	61	88	226	240	353	472	710	726	628	399	146	60	4,109
Kano, Nigeria	T	21	24	28	31	30	28	26	25	26	27	25	22	26
12°N; 476 m	P	0	T	2	8	71	119	209	311	137	14	T	0	872
Mombasa, Kenya	T	28	28	28	28	26	25	24	24	25	26	27	28	26
4°S; 55 m	P	30	14	59	192	319	100	72	69	71	86	74	76	1,163

TABLE B.6 Continued

		J	F	M	A	M	J	J	A	S	O	N	D	Yr
Tamatave, Malagasy	T	26	26	26	25	23	21	21	21	22	23	25	26	24
18°S; 5 m	P	420	442	528	403	303	300	257	208	135	91	184	259	3,530
Wadi Halfa, Sudan	T	15	16	21	26	31	32	32	32	30	28	22	17	25
22° N; 160 m	P	0	0	0	0	1	0	1	0	0	1	0	0	3
Asia														
Ankara, Turkey	T	0	1	5	11	16	20	23	23	18	13	7	2	12
40° N; 902 m	P	37	36	36	37	49	30	14	9	17	24	30	43	360
Ashkhabad, U.S.S.R.	T	2	5	9	16	23	29	31	29	24	16	8	3	16
38° N; 230 m	P	22	21	44	38	28	6	2	1	3	11	15	19	210
Baghdad, Iraq	T	10	12	16	22	28	33	35	34	31	25	17	11	23
33° N; 34 m	P	26	28	28	17	7	T	0	T	T	3	21	26	156
Bangkok, Thailand	T	26	28	29	30	30	29	28	28	28	28	27	26	28
14° N; 12 m	P	9	29	34	89	166	171	178	191	306	255	57	7	1,492
Bombay, India	T	24	25	27	29	30	29	28	27	27	28	28	26	27
19° N; 11 m	P	2	1	0	3	16	520	709	419	297	88	21	2	2,078
Inchon, Korea	T	−4	−2	3	10	15	20	24	25	21	14	7	0	11
37° N; 70 m	P	16	18	50	66	73	139	304	180	137	45	35	30	1,089
Irkutsk, U.S.S.R.	T	−21	−18	−9	2	9	15	18	15	8	1	−11	−18	−1
52° N; 485 m	P	12	8	9	15	29	83	102	99	49	20	17	15	458
Jakarta, Indonesia	T	26	26	27	27	27	27	27	27	27	27	27	27	27
6°S; 7 m	P	335	241	201	141	116	97	61	50	78	91	151	193	1,755

TABLE B.6 Continued

		J	F	M	A	M	J	J	A	S	O	N	D	Yr
Lanchow, China	T	−6	−2	5	12	17	21	23	21	16	10	2	−5	10
36° N; 1,508 m	P	1	3	8	14	34	40	66	92	55	18	4	2	338
Lhasa, Tibet (1941–48)	T	−2	1	5	8	12	17	16	16	14	9	4	0	8
30° N; 3,685 m	P	1	13	8	5	25	63	122	89	66	13	3	0	408
Manila, Philippines	T	25	26	27	29	29	28	28	27	27	27	26	25	27
15° N; 15 m	P	18	7	6	24	110	236	253	480	271	201	129	56	1,791
New Delhi, India	T	14	17	23	29	34	34	31	30	29	26	20	16	25
29° N; 216 m	P	25	22	17	7	8	65	211	173	150	31	1	5	715
Peking, China	T	−5	−2	5	14	20	25	26	25	20	13	4	−3	12
40° N; 52 m	P	4	5	8	17	35	78	243	141	58	16	11	3	623
Saigon, Viet Nam	T	26	26	28	29	28	27	27	27	27	26	26	26	27
11° N; 10 m	P	6	13	12	65	196	285	242	277	292	259	122	37	1,808
Shanghai, China	T	4	4	9	14	19	23	28	28	23	18	12	6	16
31° N; 5 m	P	46	66	77	84	107	168	139	132	160	62	53	41	1,135
Tashkent, U.S.S.R.	T	0	3	7	14	20	25	27	25	19	13	5	1	13
41° N; 428 m	P	49	51	81	58	32	12	4	3	3	23	44	57	417
Tbilisi, U.S.S.R.	T	1	3	6	12	18	21	25	24	20	14	8	3	13
42° N; 490 m	P	20	21	36	43	87	69	50	37	42	46	37	20	508
Verkhoyansk, U.S.S.R.	T	−47	−43	−30	−14	3	13	16	11	3	−14	−36	−44	−15
68° N; 137 m	P	7	5	5	4	5	25	33	30	13	11	10	7	155

TABLE B.6 Continued

Australia and New Zealand		J	F	M	A	M	J	J	A	S	O	N	D	Yr
Alice Springs, N. Terr.	T	28	27	25	20	16	12	12	14	18	23	25	28	21
24°S; 546 m	P	27	45	18	10	18	15	14	10	6	25	23	39	250
Auckland, New Zealand	T	19	19	18	16	14	12	11	11	12	14	16	18	15
37°S; 5 m	P	85	104	71	108	123	139	139	108	98	106	89	78	1,248
Brisbane, Queensland	T	25	25	24	21	18	16	15	16	18	21	22	24	20
27°S; 41 m	P	143	183	147	78	57	56	49	30	45	77	92	136	1,092
Christchurch, New Zealand	T	16	17	15	12	9	6	5	7	9	11	14	16	11
43°S; 29 m	P	47	51	52	51	81	52	53	51	48	49	42	56	633
Hobart, Tasmania	T	16	16	15	12	11	8	8	9	11	12	14	15	12
43°S; 54 m	P	42	47	52	63	51	66	47	53	53	72	58	64	668
Invercargill, New Zealand	T	13	14	12	10	8	6	5	6	8	10	11	12	10
46°S; 1 m	P	99	94	96	98	98	104	75	71	72	80	88	88	1,063
Perth, W. Australia	T	23	24	22	19	16	14	13	14	15	16	19	22	18
32°S; 60 m	P	7	12	22	52	125	192	183	135	69	54	23	15	889
Port Hedland, W. Australia	T	30	31	30	28	23	20	19	21	24	26	29	30	26
20°S; 8 m	P	56	86	36	15	45	21	15	4	1	1	1	19	299
Sydney, New South Wales	T	22	22	21	18	16	13	12	13	15	18	19	21	18
34°S; 42 m	P	104	125	129	101	115	141	94	83	72	80	77	86	1,205
Wellington, New Zealand	T	15	16	15	13	11	9	8	8	10	11	13	14	12
41°S; 119 m	P	74	104	80	90	127	123	128	116	92	116	79	95	1,224

TABLE B.6 Continued

Europe		J	F	M	A	M	J	J	A	S	O	N	D	Yr
Athens, Greece	T	9	10	11	15	20	25	28	27	24	19	15	11	18
38° N; 107 m	P	62	36	38	23	23	14	6	7	15	51	56	71	402
Berlin, Germany	T	−1	0	4	9	14	17	19	18	14	9	4	1	9
52° N; 55 m	P	45	42	31	43	49	65	72	71	47	49	48	43	605
Edinburgh, Scotland	T	4	4	5	7	10	13	15	14	12	9	6	4	8
56° N; 76 m	P	55	42	44	39	52	53	71	78	60	67	58	57	676
Kiev, U.S.S.R.	T	−6	−5	0	8	15	19	20	19	14	8	1	−3	7
50° N; 179 m	P	43	39	35	46	56	66	70	72	47	47	53	41	615
Leningrad, U.S.S.R.	T	−8	−8	−4	3	10	15	18	17	11	5	0	−4	5
60° N; 4 m	P	36	32	25	34	41	54	69	77	58	52	45	36	559
Lisbon, Portugal	T	11	12	14	16	17	20	22	22	21	18	14	12	17
39° N; 95 m	P	111	76	109	54	44	16	3	4	33	62	93	103	708
London, England	T	4	4	7	9	12	16	18	17	15	11	7	5	10
51° N; 45 m	P	51	38	36	46	46	41	51	56	46	58	63	51	583
Madrid, Spain	T	5	6	10	13	16	21	24	24	20	14	9	6	14
40° N; 660 m	P	38	34	45	44	44	27	11	14	31	53	47	48	436
Milan, Italy	T	2	4	9	14	18	22	25	24	20	14	8	3	14
45° N; 147 m	P	60	56	66	83	101	75	51	68	72	91	98	70	891
Odessa, U.S.S.R	T	−2	−2	2	8	15	20	22	22	17	11	5	0	10
46° N; 64 m	P	28	26	20	27	34	45	34	37	29	35	43	31	389
Oslo, Norway	T	−5	−4	0	5	11	15	17	16	11	6	1	−2	6
60° N; 96 m	P	49	35	26	44	44	71	84	96	83	76	69	63	740

		J	F	M	A	M	J	J	A	S	O	N	D	Yr
Rome, Italy	T	7	8	10	14	18	22	25	24	21	17	12	8	15
42°N; 18 m	P	79	69	72	67	57	39	15	25	69	126	117	99	834
Vienna, Austria	T	-1	0	5	10	15	18	20	19	16	10	5	1	10
48°N; 203 m	P	40	43	45	45	70	67	83	72	41	56	53	45	660
Warsaw, Poland	T	-2	-3	1	7	13	17	19	18	13	8	3	0	8
52°N; 107 m	P	25	28	20	32	40	60	79	47	41	31	31	37	471
North America														
Angmagssalik, Greenland	T	-7	-7	-6	-3	2	6	7	7	4	0	-3	-5	0
66°N; 36 m	P	57	81	57	55	52	45	28	70	72	96	87	75	775
Boston, Massachusetts	T	-3	-3	2	8	14	18	22	21	17	12	6	-1	9
42°N; 198 m	P	114	95	115	102	88	95	83	103	100	95	115	101	1,207
Chicago, Illinois	T	-3	-2	2	10	16	22	24	24	19	13	4	-2	10
42°N; 187 m	P	47	41	70	77	95	103	86	80	69	71	56	48	843
Churchill, Manitoba	T	-28	-26	-20	-10	-2	6	12	12	6	-1	-12	-22	-7
59°N; 35 m	P	17	17	21	36	40	48	39	60	52	42	46	25	443
Coppermine, N.W.T.	T	-29	-30	-26	-17	-6	3	9	8	3	-7	-20	-26	-11
68°N; 9 m	P	13	8	15	14	12	20	34	44	29	27	17	13	246
Dallas, Texas	T	8	10	14	18	23	27	29	29	26	20	14	9	19
33°N; 156 m	P	63	61	84	107	114	97	71	76	69	71	69	64	946
Edmonton, Alberta	T	-14	-12	-6	4	11	14	17	16	11	5	-4	-10	3
53°N; 676 m	P	24	20	21	28	47	80	85	65	34	23	22	25	474
Gander, Newfoundland	T	-6	-6	-4	1	6	12	17	16	12	6	2	-4	4
49°N; 147 m	P	83	88	81	69	59	76	84	101	86	97	107	87	1,018

TABLE 5.6 Continued

		J	F	M	A	M	J	J	A	S	O	N	D	Yr
Havana, Cuba	T	22	22	23	25	26	27	28	28	28	26	24	23	25
23° N; 24 m	P	71	46	46	58	119	165	124	135	150	173	79	58	1,224
Mazatlan, Mexico	T	20	20	20	22	24	27	28	28	28	27	24	21	24
23° N; 78 m	P	12	8	3	0	1	34	174	215	250	63	17	27	805
Mexico City, Mexico	T	13	15	17	18	19	18	17	17	17	16	14	13	16
19° N; 2,259 m	P	4	5	11	18	46	100	116	114	102	37	13	5	571
Miami, Florida	T	19	20	21	23	25	27	28	28	27	25	22	20	24
26° N; 4 m	P	52	47	58	99	164	187	171	177	241	209	72	42	1,518
Nome, Alaska	T	−15	−15	−13	−6	2	8	10	9	6	−1	−9	−14	−3
64° N; 339 m	P	26	24	22	20	18	24	58	97	68	43	29	25	454
Phoenix, Arizona	T	10	12	16	20	25	30	33	32	29	22	15	11	21
33° N; 339 m	P	19	22	17	8	3	2	20	28	19	12	12	22	183
Salt Lake City, Utah	T	−2	1	5	10	15	19	25	24	18	12	3	0	11
41° N; 1,287 m	P	34	30	40	45	36	25	15	22	13	29	33	31	353
Spokane, Washington	T	−4	−1	3	8	13	16	21	20	16	9	2	−1	8
48° N; 721 m	P	62	47	38	23	31	38	10	10	19	40	57	62	437
Topeka, Kansas	T	−2	1	5	12	18	24	27	26	21	15	6	1	13
39° N; 269 m	P	24	44	58	72	149	119	141	82	75	44	19	25	854
Yakutat, Alaska	T	−3	−2	0	3	7	10	12	12	10	6	1	−2	4
60° N; 9 m	P	276	208	221	184	203	129	214	277	420	498	407	312	3,348
Yuma, Arizona	T	13	15	19	22	26	31	35	34	31	25	18	14	24
33° N; 62 m	P	10	9	6	2	T	T	6	13	10	10	3	8	77

TABLE B.6 Continued

South America		J	F	M	A	M	J	J	A	S	O	N	D	Yr
Antofagasta, Chile (1951–60)	T	20	20	19	16	15	14	13	14	14	15	17	18	16
23°S; 122 m	P	0	0	0	0	0	0	0.3	0.1	0	0	0	0	0.4
Bogota, Colombia (1951–60)	T	13	14	14	14	14	14	13	13	13	14	14	14	14
5°N; 2,556 m	P	38	48	66	96	103	60	48	38	53	158	140	93	942
Caracas, Venezuela (1951–60)	T	19	20	21	22	22	22	21	22	22	22	21	20	21
10°N; 1,042 m	P	22	28	11	43	92	121	107	104	107	110	73	44	862
Manaus, Brazil	T	26	26	26	26	26	27	27	28	28	28	27	27	27
3°S; 44 m	P	278	278	300	287	193	99	61	41	62	112	165	220	2,095
Maracaibo, Venezuela	T	26	27	27	28	28	29	29	29	29	28	28	27	28
11°N; 40 m	P	3	2	3	27	73	43	28	42	40	99	22	5	387
Montevideo, Uruguay	Y	23	22	20	17	14	11	10	11	13	15	18	21	16
35°S; 22 m	P	83	74	104	102	91	88	73	87	84	73	79	77	1,014
Rio de Janeiro, Brazil	T	26	26	26	24	22	21	21	21	22	22	23	24	23
23°S; 27 m	P	136	137	133	116	73	44	43	43	53	74	97	127	1,074
Sao Paulo, Brazil	T	22	22	21	19	17	16	15	16	17	19	19	20	18
24°S; 795 m	P	224	209	166	69	53	42	31	42	66	113	118	155	1,287
Ushuaia, Argentina	T	9	9	8	6	3	2	2	2	4	6	7	8	6
55°S; 6 m	P	58	50	57	46	48	45	47	50	38	36	50	49	574
Valdivia, Chile (1951–60)	T	16	16	15	12	10	8	8	8	9	12	14	16	12
40°S; 13 m	P	106	50	96	178	437	371	452	318	225	98	82	72	2,486

TABLE B.6 Continued

Islands		J	F	M	A	M	J	J	A	S	O	N	D	Yr
Apia, Western Samoa	T	27	27	27	27	27	26	26	26	26	26	27	27	27
14°S, 172°W; 2 m	P	424	364	352	214	186	130	115	111	147	221	279	385	2,928
Ascension, South Atlantic	T	26	27	28	28	27	26	25	25	24	24	25	26	26
8°S, 14°W; 17 m (1951–60)	P	3	3	12	20	14	19	14	12	8	9	4	5	122
Canton, Phoenix Islands	T	28	28	28	29	29	29	29	29	29	29	29	28	29
3°S, 172°W; 3 m	P	66	54	63	92	110	67	66	64	31	28	41	65	748
Guam, Marianas	T	26	26	26	27	27	27	26	26	26	26	26	26	26
14°N, 145°E; 110 m	P	118	89	67	77	106	149	228	326	339	333	261	155	2,249
Honolulu, Hawaii	T	23	22	23	23	24	26	26	26	26	26	24	23	24
19°N, 158°W; 12 m	P	96	84	73	33	25	8	11	23	25	47	55	76	556
Jan Mayen, North Atlantic	T	-4	-5	-5	-3	0	2	5	6	4	1	-1	-3	0
71°N, 8°W; 39 m	P	79	54	63	58	23	28	36	61	83	93	82	75	735
Laurie, South Orkneys	T	0	0	-1	-3	-7	-10	-11	-10	-7	-4	-2	-1	-5
61°S, 45°W; 4 m	P	35	39	48	41	32	26	32	32	29	29	32	27	402
Macquarie, South Pacific	T	7	7	6	5	4	3	3	3	3	4	4	6	5
54°S, 159°E; 6 m	P	99	81	104	94	81	81	68	68	68	73	64	72	952
Mauritius, Indian Ocean	T	26	26	26	24	22	21	20	20	21	22	24	26	23
20°S, 58°E; 55 m	P	207	195	218	141	97	67	57	59	37	39	51	115	1,283
Noumea, New Caledonia	T	26	26	25	23	22	21	19	20	21	22	24	25	23
22°S, 166°E; 72 m	P	117	94	175	124	93	89	84	70	53	52	47	85	1,083
Sal, Cape Verde Islands	T	21	21	21	21	22	23	24	26	27	26	24	22	23
16°N, 23°W; 55 m	P	2	2	1	0	0	0	18	20	27	15	10	13	108
Tristan da Cunha, South	T	17	18	17	16	14	13	12	12	12	13	14	16	15
Atlantic 37°S, 12°W; 23 m	P	103	88	88	153	144	187	147	193	129	139	117	167	1,655

Bibliography

BOOKS

ATKINSON, BRUCE W., *The Weather Business.* Garden City, N.Y.: Doubleday & Company, Inc., 1969.

BARRY, R. G., and R. J. CHORLEY, *Atmosphere, Weather, and Climate.* New York: Holt, Rinehart and Winston, Inc., 1970.

BLAIR, THOMAS A., and ROBERT C. FITE, *Weather Elements,* 5th ed. Englewood Cliffs, N.J.: Prentice-Hall, Inc., 1965.

CHANG, JEN-HU, *Atmospheric Circulation Systems and Climates.* Honolulu: Oriental Publishing Company, 1972.

COLE, FRANKLYN W., *Introduction to Meteorology.* New York: John Wiley & Sons, Inc., 1970.

CROWE, P. R., *Concepts in Climatology.* New York: St. Martin's Press, Inc., 1971.

DAY, JOHN A., and GILBERT L. STERNES, *Climate and Weather.* Reading, Mass.: Addison-Wesley Publishing Company, Inc., 1970.

FLOHN, HERMANN, *Climate and Weather.* New York: McGraw-Hill Book Company, 1969.

Geiger, Rudolf, *The Climate Near the Ground,* 4th ed. Cambridge, Mass.: Harvard University Press, 1965.

Griffiths, John F., *Applied Climatology*. New York: Oxford University Press, 1966.

Hare, F. Kenneth, *The Restless Atmosphere,* 3rd ed. London: Hutchinson & Co. (Publishers), Ltd., 1961.

Haurwitz, Bernhard, and James M. Austin, *Climatology*. New York: McGraw-Hill Book Company, 1944.

Huschke, Ralph E., *Glossary of Meteorology*. Boston: American Meteorological Society, 1959.

Kendrew, W. G., *The Climates of the Continents,* 5th ed. London: Oxford University Press, 1961.

Landsberg, Helmut, *Physical Climatology,* 2nd ed. Dubois, Pa.: Gray Printing Company, Inc., 1958.

List, Robert J., *Smithsonian Meteorological Tables,* Sixth Revised Edition (Smithsonian Misc. Collections, Vol. 114). Washington: Smithsonian Institution, 1966.

Longley, Richmond W., *Elements of Meteorology*. New York: John Wiley & Sons, Inc., 1970.

Lowry, William P., *Weather and Life*. New York: Academic Press, 1969.

McIntosh, D. H., *Meteorological Glossary,* 5th ed. London: H. M. Stationery Office, 1972.

Miller, Albert, and Jack C. Thompson, *Elements of Meteorology*. Columbus, O.: Charles E. Merrill Publishing Co., 1970.

Neiburger, Morris, James G. Edinger, and William D. Bonner, *Understanding Our Atmospheric Environment*. San Francisco: W. H. Freeman and Company, 1973.

Petterssen, Sverre, *Introduction to Meteorology,* 3rd ed. New York: McGraw-Hill Book Company, 1969.

Riehl, Herbert, *Introduction to the Atmosphere,* 2nd ed. New York: McGraw-Hill Book Company, 1972.

Rumney, George R., *Climatology and the World's Climates*. New York: The Macmillan Company, 1968.

Sellers, William D., *Physical Climatology*. Chicago: University of Chicago Press, 1965.

Shaw, Robert H., ed., *Ground Level Climatology,* Pub. No. 86. Washington: American Association for the Advancement of Science, 1967.

Stringer, E. T., *Foundations of Climatology*. San Francisco: W. H. Freeman and Company, 1972.

———, *Techniques of Climatology*. San Francisco: W. H. Freeman and Company, 1972.

TREWARTHA, GLENN T., *An Introduction to Climate,* 4th ed. New York: McGraw-Hill Book Company, 1968.

United States Department of Commerce, Weather Bureau, *World Weather Records, 1941–50.* Washington: Government Printing Office, 1959.

———, *World Weather Records, 1951–60,* 6 vols. Washington: Government Printing Office, 1965–68.

WILLETT, HURD C., and FREDERICK SANDERS, *Descriptive Meteorology,* 2nd ed. New York: Academic Press Inc., 1959.

PERIODICALS

Agricultural Meteorology. Amsterdam: Elsevier Publishing Company, quarterly.

Atmosphere, the Journal of the Canadian Meteorological Society. Toronto: University of Toronto Press, quarterly.

Bulletin of the American Meteorological Society. Boston: American Meteorological Society, monthly.

Canadian Weather Review. Downsview, Ont.: Atmospheric Environment Service, monthly.

Climatological Data, National Summary. Washington, D.C.: Department of Commerce, Environmental Data Service, monthly.

Climatological Data of the United States (by sections). Washington, D.C.: Department of Commerce, Environmental Data Service, monthly.

Daily Weather Maps, Weekly Series. Washington, D.C.: Department of Commerce, Environmental Data Service, weekly.

Daily Weather Report. London: H. M. Stationery Office, daily.

Journal of Applied Meteorology. Boston: American Meteorological Society, bimonthly.

Journal of Atmospheric Sciences. Boston: American Meteorological Society, bimonthly.

Local Climatological Data. Washington: Department of Commerce, Environmental Data Service, monthly and annual summary.

Marine Observer. London: H. M. Stationery Office, quarterly.

Mariner's Weather Log. Washington, D.C.: Department of Commerce, Environmental Data Service, bimonthly.

Meteorological Magazine. London: H. M. Stationery Office, monthly.

Monthly Climatic Data for the World. Washington, D.C.: Department of Commerce, monthly.

Monthly Weather Report. London: H. M. Stationery Office, monthly.

Monthly Weather Review. Washington, D.C.: Department of Commerce, National Oceanic and Atmospheric Administration (Boston: American Meteorological Society, beginning with Vol. 102, January 1974), monthly.

Quarterly Journal of the Royal Meteorological Society. London: Royal Meteorological Society, quarterly.

Weather. London: Royal Meteorological Society, monthly.

Weatherwise. Boston: American Meteorological Society, bimonthly.

World Meteorological Organization Bulletin. Geneva: World Meteorological Organization, quarterly.

Index

A

Ablation, 255, 373
Absolute humidity, 36
Absolute scale, 23
Absolute zero, 23
Accidents, 316, 328
Acclimatization, 363–65
  definition, 363
Actual lapse rate, 32
Adelaide, Australia, 177
Adiabatic cooling, 43–44
Adiabatic heating, 127–29
Adiabatic processes, 43–44
Adriatic Coast, 176
Adriatic Sea, 87
Advection fog, 55, 402–3
Aerial reconnaissance, 130
Aeroallergens, 360
Aerology, 4
Aeronomy, 4
Aerosol, 57
  urban, 385–88
Afghanistan, 170
Africa, 92, 97, 99, 169–70, 175–76, 178, 212, 229, 300 (*See also* names of countries)
Agata, Siberia, 72
Agriculture:
  climate and, 265–301
  irrigation and, 284–86
Agroclimatic analogs, 301
Agung, 380
Ahaggar Mountains, 170, 212
Ahlmann, Hans, 373–74
Air, 5 (*See also* Atmosphere)
  advection of, 44, 55
  density of, 33, 69, 71
  Greek concept of, 6–7
  temperature (*See* Temperature)
  vertical movements of, 76
  water vapor and, 7
Air conditioning, 340
Aircraft engine design, 305
Aircraft icing, 309–10
Air currents:
  and air pollution, 324
  altimeters and, 307
Air drainage, and frost, 270
Air glow, 11
Air masses, 100–108, 247
  aerial reconnaissance in, 130–32
  analysis of, 101–3
  classification, 106–8

Air masses (*Cont.*)
  conditional instability of, 106
  continental, 108
  definition, 100
  designation, 106–8
  and fronts, 109–10
  in humid continental climates, 193
  in humid continental cool summer climate, 197
  in humid subtropics, 178–79
  identification and analysis, 101–3
  instability of, 104–6
  in marine climate, 183
  maritime, 108, 183, 247
  neutral equilibrium of, 104
  observations, 130–32
  in polar climates, 209
  properties of, 100
  in rainy tropics, 160
  source regions, 100–101, 106–8
  special effects in, 127–30
  stability of, 104–6
  in taiga climate, 203
  and tornadoes, 127
  in tropical arid and semiarid climates, 168–69
  tropical and polar, 174
  in tundra climate, 206
  in wet-and-dry tropics, 167
Air mass weather, 110
Airplanes, frost prevention and, 272
Air pollution, 322–26
  and air conditioning, 346
  and health, 359–60
  and housing sites, 331–32
  and industry, 322–25
Air pollution index, 360
Air pollution potential, 325–26
Airport runways, and winds, 304
Airport sites, 304
Air Quality Act, 360
Alabama, 193
Alaska, 182, 202, 207–9, 277, 396
Alaska peas, 279
Albedo, 13
  and climatic change, 381
  of dust, 390
  and modification of climate, 393, 402
  table, 14
Alberta, 197
Aleutian Islands, 208
Algae, 222
Alice Springs, Australia, 152
Alkanols, 395, 402
Allahabad, India, 65, 179
Allergies, 360, 362
Alpine meadows, 234
Alpine soils, 241
Alps, 88, 127, 212, 375
Altimeters, 71, 307–8
Altitude:
  acclimatization to, 364–65
  climate and, 212–16
  and insolation, 213–14
  of snowline, 215–16
  temperature and, 9, 32–33, 212–13
  vegetation and, 233–35
  winds and, 79–80, 213
Altocumulus, 47–48
Altocumulus lenticularis, 48
Altostratus, 47
Amazon Basin, 156
American Society of Heating, Refrigerating and Air-Conditioning Engineers, 341
Amundsen-Scott Station, 210
Analogs, 136–37, 301
Andes Mountains, 63, 170, 212, 296
Anemometers, 81–83
Aneroid barometer, 71
Angmagssalik, Greenland, 208
Animals:
  and climate, 293–97
  in forests, 227
  introduction of, 300–301
  and temperature, 294
Annual march of temperature, 26
Annual range of temperature, 26
Annual regime of precipitation, 65
Antarctica, 206, 209–12
Antarctic air masses, 108
Anticyclones, 116–17, 204
Aphelion, 16
Applied classification, 142–43
Arabia, 169
Archangel, U.S.S.R., 204
Architecture, climate and, 335–40
Arctic air masses, 108, 206
Arctic Ocean, 206–10
Argentina, 63, 88, 178, 188–89, 191, 230
Argon, 7
Arica, Chile, 172
Arid climates:
  mid-latitude, 188–92
  thunderstorms in, 256
  tropical, 168–72
Arizona, University of, 373
Artificial nucleation, 398
Asia, 162–63, 166–67, 169, 189, 193, 196, 198, 200, 212–13, 228, 256 (*See also* names of countries)
Asia Minor, 229
Aswan Dam, 394
Atacama, 169
Athens, Greece, 185–86
Atlantic Ocean, 208–9

Atlas Mountains, 63, 170
Atmometers, 40
Atmosphere, 3–6 (*See also* Air)
  and climatic change, 380–81
  composition of, 6–8
  heating and cooling of, 12–14
  mass of, 5
  nature of, 5–6
  origin of, 6
  standard, 72
  transparency of, 16–17
  upper, 9–11
  vertical divisions, 8–10
  water budget of, 35
Atmospheric circulation, 88–90
  seasonal changes in, 93–96
Atmospheric dust, 380–81, 390
Atmospheric heat budget, 12–15, 20
Atmospheric moisture, 35–37, 246
Atmospheric pollution, 322–26
  and health, 359–60
Atmospheric pressure, 68
  altitude and, 72, 213, 307–8
  animals and, 296
  health and, 356, 364
  high, 73
  horizontal distribution of, 72–75
  low, 73
  measurement, 69–71
  sea level, 69, 72
  seasonal changes, 73
  world maps, 74–75
Atmospheric turbulence, 308–9
ATS photograph, 132
Attic venting, 344
Aurora australis, 10
Aurora borealis, 10, 208
Australia, 129, 164, 169, 175, 178, 182, 228, 365
Australian Bight, 188
Australian desert, 229
Austria, 127
Avalanches, snow, 262–63
Avalanche winds, 263
Aviation, climatology and, 304
Aviation weather reports, 304–5
Azizia, Libya, 170
Azonal soils, 238
Aztecs, 375

B

Bahia Felix, Chile, 186
Baiu rains, 180
Balkash, U.S.S.R., 189, 191
Barographs, 71
Barometers, 69–71
Barometric fog, 56
Barometric pressure (*See also* Atmospheric pressure):
  decrease with altitude, 72, 213
  at warm front, 111
Barrow, Alaska, 207
Bay of Bengal, 182
Beaufort, Admiral Francis, 80
Beaufort scale, 81
Belgium, 360
Benguela Current, 31, 169
Bering Strait, 208–9, 396
Bibliography, 427–30
Biosphere, 3
Birmingham, England, 185–86
Black frosts, 269
Black Sea, 178, 229
"Blind landing" equipment, 304
Blizzards, 113, 192, 196, 212
Blood circulation, 356, 363–64
Bolivia, 164
Bora, 87, 176
Boreal forest climate, 202
Borneo, 212
Bourdon tube, 24
Brazil, 160, 164
Breeding of livestock, 297, 300
Bridge construction, 327
British Columbia, 185
British Isles, 182, 374
Broadleaf forests, 228
Bucharest, Romania, 194, 197
Budyko, M. I., 252
Buenos Aires, Argentina, 179–80
Building orientation, 333–34
Building sites, climate and, 330–33
Buran, 113
Burma, 160
Bush tundra, 231
Buys-Ballot's rule, 79

C

Calcium chloride, 403
California, 175–76, 180, 182, 228, 235, 273
Cambrian period, 6, 369
Campbell-Stokes recorder, 18–19
Campos, 229
Canada, 188, 206, 228, 256–57, 373
Cape Race, Newfoundland, 199
Cape São Roque, Brazil, 160, 170
Carbon dioxide, 7
  and climatic change, 381–82, 389
Carbon-14 method of dating, 371
Carbon monoxide, 359
Carburetor icing, 310
Caribbean Sea, 159
Cascade Mountains, 212, 216
Cattle, 295

Cayenne, French Guiana, 162
Ceiling:
  of clouds, 52
  definition, 52
Ceiling balloons, 52
Ceiling light, 52
Ceilometer, 52
Celsius scale, 21
Centigrade scale, 21
Central America, 156, 164
Chaparral, 228
Charleston, South Carolina, 179
Charts:
  synoptic, 132–34
  upper-air, 134
Cherrapunji, India, 59, 66, 182
"Cheyenne fog," 56
Chile, 63, 169, 175–76, 182, 186
Chilling:
  and illness, 361
  of plants, 267–68, 286
China, 178, 188, 193, 196
Chinook, 88, 127–29, 197
Circulation of blood, 356
Cirrocumulus, 47
Cirrostratus, 47
Cirrus, 45–47
Cities, and air pollution, 304, 331, 385–87
City climates, 385–89
Clean Air Act, 360
Clear air turbulence, 309
Clear ice, 309–10
Clerestories, 335
Cliff dwellers, 375
Climate:
  and acclimatization, 363–65
  arid, 168–72, 188–92
  classification of, 141–52
  and construction, 326–28
  definition, 3
  and disease, 361–63
  drought and, 281–84
  in dry summer subtropics, 174–77
  elements of, 4
  and floods, 255–58
  and food, 347–48, 356–57
  forecasts of, 382
  and forestry, 288–92
  and health, 355–59
  highland, 212–16
  house sites and, 330–33
  and human comfort, 347–66
  humid continental cool summer, 196–200
  humid continental warm summer, 192–96
  in humid subtropics, 178–82
  and local relief, 213
  marine, 182–88
  in monsoon tropics, 160–64
  nutrition and, 356–57
  perception of, 365–66
  in rainy tropics, 156–60
  during recorded history, 375–76
  semiarid, 168–72, 188–92
  soils and, 235–44
  subarctic, 202
  taiga, 202–6
  tundra, 206–9
  vegetation and, 221–35
  in wet-and-dry tropics, 164–68
Climate conditioning, 333
Climate modification, 384–404
  implications of, 404
Climatic adaptation, 297
Climatic analogs, 301
Climatic change, 368–82
  astronomical theories of, 380
  carbon dioxide and, 381–82, 389
  and continental drift, 381
  glaciers and, 373–75
  theories of, 378–82
Climatic cycles, 377–78
Climatic data, supplementary, 416–23
Climatic fluctuations, 373, 377
  and forests, 292
Climatic optima, 223–24
Climatic region, 141
Climatic trends, recent, 376–77
Climatological benchmark network, 376
Climatologists, 145, 148, 152
Climatology:
  applied, 4–5, 144, 303–4
  definition of, 4
  physical, 4
  regional, 4–5, 140
Climax vegetation, 226
Clinometer, 52
Clothing, climate and, 353–54
Cloudbursts, 123, 236, 243
Cloud ceiling, 52
Cloud modification, 397–400
Cloud observations, 51–53
Clouds, 45–53
  albedo of, 13
  classification, 45–51
  direction, 52–53
  formation, 45
  height, 52
  speed of movement, 52
  supercooled, 398–99
Cloud seeding, 398–401
  and forest-fire prevention, 292
Cloud types, 51

Coalescence:
  of fog particles, 403
  of raindrops, 57, 397–400
Cobalt-60 isotope, 261
Coffeyville, Kansas, 59
Cold fronts, 109–10, 112–13
College Park, Maryland, 279
Colorado, 189, 191, 214, 280, 375
Comfort chart, 341, 343
Comfort zone, 341, 343, 350
Communications, public, 320–21
Condensation, 41–44
  latent heat of, 42, 44, 127
Condensation nuclei, 8, 42–43
Condensation temperature, 42
Conduction, 14
Congo Basin, 156
Coniferous forests, 227–28
Constant-level charts, 103
Constant-pressure analyses, 103
Constant-pressure charts, 103
Construction and weather, 326–28
Continental air masses, 108, 190, 197, 200
Continental drift, and climatic change, 381
Convection, 15
  thermal, 9
Convectional precipitation, 59
Convective motion, 9
Convergence, 43–44
Conversion equivalents, 409–10
Cooling:
  adiabatic, 43–44
  by evaporation, 270, 342
  of houses, 345–46
  of livestock shelters, 296
Cooling degree-days, 345–46
Cooling power, 350–52
Coquilhatville, Zaire, 157
Coriolis effect, 77–78, 89, 97
Corn Belt, 193
Coromandel Coast, 165
Corona, 48
Corpuscular energy, 379
Cranberry bogs, and frost, 273
Crops:
  climatic factors and, 266–87
  hail damage and, 280
  and irrigation, 284–86
  temperature and, 266–71
  tropical, 298
  and wind, 286–88
Crossbreeding for climatic adaptation, 297, 300
Cuiaba, Brazil, 165
Cumulonimbus, 50–51
Cumulus, 49–51
Cwilong, Bohdan M., 397
Cycles:
  climatic, 377–78
  sunspot, 378–79
Cyclones, 196
  extra-tropical, 108–11
  tropical, 117–21
Cyclonic storms, in marine climates, 187
Czechoslovakia, 197

D

Daily mean temperature, 25
Daily Weather Map, 135
Damp haze, 54
Danube Basin, 193
Darwin, Australia, 168
Dawson, Yukon Territory, 204
Daylight:
  duration of, 16–18
  influence on animals, 295
  latitude and, 198, 203–4
De Candolle, Alphonse, 145
Deccan Plateau, 169
Decomposition, 236
Degree-days:
  cooling, 345–46
  growing, 277
  heating, 344–45
Dendrochronology, 373
Denver, Colorado, 189
Desert climate, 169, 190
Deserts, 169, 171, 230
  soils of, 241
  vegetation in, 190, 230
Desert tundra, 231
Des Moines, Iowa, 193
Dew, 42, 44
Dew point, 42, 270
Dew point hygrometer, 39
Diabatic processes, 43–44
Diamond dust, 55
Dickinson, North Dakota, 190–91
Diseases:
  of animals, 299–300
  and climate, 361–63
Disintegration, 235
Diurnal range, 25–26
Diurnal variation, of temperature, 387
Doctor, 87
Doldrums, 73, 88
Donora, Pennsylvania, 360
Douglass, A. E., 373
Downdrafts, 307–8
Dresden, Ohio, 259
Drifts, ocean, 97
Drizzle, 58
Dropsondes, 131

Drought, 281–84
  and irrigation, 284–86
  and monsoons, 283
  permanent, 281, 283
  physiological, 268
  precipitation variability and, 281, 283
  seasonal, 281, 283
  and temperature, 281
Dry adiabatic rate, 44
Dry convective cooling power, 351
Dry farming, 284
Dry ice, 43
  and cloud modification, 397–98
  and fog dispersal, 402
Dry summer subtropics, 174–77
  and floods, 256
Dundee, Scotland, 327
Durban, South Africa, 181
Dust:
  atmospheric, 8, 380, 390
  and condensation, 8, 42
  in ice layers, 390
  and radiation, 8
"Dust Bowl," 244
Dust-devils, 129–30
Dust storms, 129
Dyne, 69

E

East Africa, 170
Easterly jet stream, 93
Easterly waves, 117–18
Eastern Highlands, of Africa, 212
Eastport, Maine, 199
Ecoclimate, 144
Ecuador, 214
Eddies, 60, 76, 78
Edmonton, Alberta, 198
Effective temperature, 341, 349–50
Egypt, 87
Eismitte, Greenland, 210
Electrical storms, 130
Electric power lines, 318–20
El Golea, Algeria, 170
El Niño, 172
Eluviation, 236
England, 185
Equatorial air masses, 107–8, 160
Equatorial convergence, 88
Equatorial low, 73, 88
Equinoxes, precession of, 380
Equivalent wind chill temperature, 351
Eskimos, 353
Etesian winds, 88
Eurasia, 202, 206 (*See also* country names)
Europe, 182, 185, 196, 228, 365, 375 (*See also* country names)
Evaporation, 39–41
  and cooling of human body, 348–49
  of fog, 402–3
  measurement of, 40
Evaporation suppressors, 268
Evaporimeters, 40
Evaporites, 241, 371
Evapotranspiration, 40–41, 286–87
  and wind, 287
Evapotranspirometers, 251–52

F

Facsimile recorder, 136
Fahrenheit scale, 21
Fairbanks, Alaska, 203, 277–78
Fans, for frost prevention, 272–73
Federal Aviation Agency, 305
Fernandina, 380
FIDO, 402
Field capacity, of soils, 252–53
Flash floods, 255–56
Flight paths, and upper winds, 306
Floods, 254–60
  climatic causes of, 255–58
  in dry summer subtropics, 256
  forecasting of, 258–62
  and frost prevention, 273
  in humid continental climates, 256
  in marine climates, 256
  in United States, 256
Florida, 256, 359
Florida Strait, 397
Flow patterns, of winds, 94–96
Flying, in storms, 310–11
Flying weather, 306–8
Foehn, 88, 127–28
Foehn effect, 127–28
Foehn wind, 88, 127–28, 216, 256
Fog, 42–44, 53–56
  and air pollution, 323
  in dry summer subtropics, 176–77
  formation, 53–56
  in highland climates, 215
  in humid continental cool summer climate, 200
  in humid continental warm summer climate, 196
  in marine climate, 188
  and ocean currents, 99
  supercooled, 402
  in taiga climate, 206
  in tropical arid climate, 169, 172
  in tundra climate, 209
  types of, 54
"Fog brooms," 403
Fog dispersal, 402–4
Fog drip, 54

Fog Investigation and Dispersal Operation (FIDO), 402
Food, and climate, 347–48, 356–57
Forecasting climate, 382
Forecasting weather, 134–37
Forest fires, 292–93, 401
Forest-fire weather, 292–93
Forestry and climate, 288–93
Forests, 226–28
  and climatic fluctuations, 292
  and precipitation, 289–90
  and temperature, 288–89
Fossils, 369, 374
Fractostratus, 49
France, 87, 176–77, 184
Freetown, Sierra Leone, 162
Freezes, 268–71, 273
Freezing, 42–43
Freezing rain, 58, 196
French Guiana, 162, 164
Frontal fog, 54–55
Frontal inversion, 33
Frontal lifting, 44
Frontal precipitation, 59–60
Fronts, 33, 109–16
  and aviation, 306, 308, 310–11
  cold, 110, 112–13
  mountain barriers and, 115–16
  occluded, 113–14
  polar, 109
  upper, 114, 116
  warm, 111–16
Frostbite, 294, 356
Frost-free season, 26–27, 193, 197, 203, 274
Frost-heave, 317
Frosts, 43
  advection, 269
  air mass, 269
  and crops, 269–71
  damage by, 269
  in dry summer subtropics, 176
  in humid subtropics, 180
  prevention of, 271–74
  radiation, 269–70
  in tundra climate, 208
Frost warnings, 273–74
Fuel moisture indicator, 293
Fusion, latent heat of, 43

G

Galerias, 229
Gales, 188, 200, 317, 327, 338
Garua, 172
Gas demand, 319–20
Geiger, Rudolf, 145
General atmospheric circulation, 88–90
  seasonal changes in, 93–96
Geochronology, 370–74
Georgetown, Guyana, 164
Georgia, 402
Germany, 127
GHOST balloons, 102
Gibraltar, Strait of, 397
Glaciers, and climatic change, 373–75
Glaciology, 373–75
Glaze, 58
Gliders, 306–7
Globe temperature, 342
Godthaab, Greenland, 208
Gondwanaland, 369
Gradient wind, 78–79
Grand Banks, 200
Grasslands, 228–30
Grass tundra, 231
Gravity winds, 87 (*See also* Katabatic winds)
Great Lakes, 159, 267, 313–14
Great Lakes States, 126
Great Plains, 188–89, 191–94, 230, 284, 395
Greece, 185–86
Greenhouse effect, 14, 381
Greenland, 209–11, 374–75, 377
Ground fog, 55, 402
Ground water, 252–54
Growing season, 274–75
Guinea Coast, 87, 160, 162
Gulf Coast, of United States, 180
Gulf of Mexico, 88
Gulf States, 126
Gulf Stream, 63, 183, 200, 397
Gusts, 78, 82
Guyana, 164

H

Hail, 56, 58–59, 280–81
  formation, 58
Hail damage, to houses, 340
Hail insurance, 280
Hailstorms, 192, 195
Hail suppression, 401
Hair hygrometers, 38
Haiti, 161
Halos, 47–48
Harbin, Manchuria, 193, 198
Harmattan, 87
Haze, 42
  damp, 54
Headwinds, 306, 311
Heat budget, 210
  albedo and, 381
  of body, 348–49
  equation, 348
Heat cramps, 355
Heat exhaustion, 355

Heating:
  adiabatic, 127–29
  of homes, 341–45
  of livestock shelters, 296
Heating degree-days, 344–45
Heating systems, 341–42
Heat island, 387–89
Heat of fusion, 43
Heat stroke, 355
Helium, 6–7
Highland climates, 212–16
High Plains, 56
Highway accidents, 316
Highway maintenance, weather and, 317–18
Highways, weather and, 316–18
Highway safety, 316–18
Himalayas, 97, 162, 212–13
Hokitika, New Zealand, 184
Hokkaido, 197, 199
Hong Kong, 179
Honshu, 193
Horizons, of soils, 237
Horse latitudes, 88
Hothouses, 268
House design, and climate, 333–40
House orientation, 333–34
House sites, 330–33
Housing and climate, 330–46
Humboldt Current, 169
Humid continental climates, 192–200
  floods in, 256
Humid continental cool summer climate, 196–200
Humid continental warm summer climate, 192–96
Humidity, 35–37
  absolute, 36
  and construction materials, 327
  and forest-fire danger, 292–93
  influences, on animals, 296
  measurements, 37–39
  relative, 37–38
  specific, 35–36
Humidity gauges, 38–39
Humid subtropics, 178–82
Hungary, 193
Hurricanes, 118–21
  damage by, 121
  "eye" of, 120
  in humid continental cool summer climate, 200
  in humid subtropics, 180
  modification of, 401–22
  naming of, 121
  and navigation, 314
  origin of, 117–19
  precipitation in, 120
  winds in, 119–20
Hydrocarbons, 359
Hydroelectricity, 318
Hydrogen, 6–7
Hydrographs, 259
Hydrologic cycle, 247–48
Hydrosphere, 3
Hygrograph, 38
Hygrometers, 38–39
Hygrophytes, 222
Hygroscopic particles, 42
Hyperthermia, 355
Hypothermia, 356

I

Ice, water transport and, 313–14
Ice ablation, 255, 373–74
Icecaps, 209–11, 368, 374
Ice crystals, 43
Ice fog, 55
Iceland, 206
Ice nuclei, 43, 397–99
Ice pellets, 58
Ice storm, 58
Icicles, 340
Icing, aircraft, 309–10
Illinois, 125
India, 65, 166–67, 169–70, 182, 229
Indiana, 125
Indian Ocean, 162, 170
"Indian summer," 196
Indonesia, 156
Industry, and weather, 321–28
Infrared hygrometer, 38
Infrared radiation, 12, 381
Insects, and weather, 297–300
Insolation:
  definition, 12
  duration of, table, 18
  in highland climates, 213
  house orientation and, 334
  house sites and, 332
  measurement, 18–20
  variability, 15–18
  in wet-and-dry tropics, 165
  world distribution, 20–21
Insulation, 339, 353–54
  footwear and, 354
  in houses, 336, 339–40
Intensity, of precipitation, 65–66
Intertropical convergence zone, 88, 160
  and locust swarms, 298–99
Intrazonal soils, 238
Ionosphere, 10–11
Iowa, 194
Iquique, Chile, 172
Iran, 170

Manchuria, 193, 197–98, 229
Manitoba, 197, 283
Manufacturing, and weather, 321–22
Maquis, 228
"Mare's tails," 111
Mariana Islands, 72
Marine climate, 182–88
  floods in, 256
Marine forecasts, 315
Maritime air masses, 108, 175, 178, 183, 193, 199, 247, 255
Maritime Provinces, 197, 199–200
Marquis wheat, 277–78
Marseille, France, 175, 177
Maryland, 279
Massachusetts, 398
Mayon, 380
Mean monthly temperature, 26
Mean radiant temperature, 342
Mediterranean Basin, 175
Mediterranean climate, 175
Mediterranean coast, 87
Mediterranean scrub forest, 227–28
Melbourne, Australia, 183
Melting, 42, 254–56
  and modification of climate, 393, 395
Memphis, Tennessee, 149, 179–80, 250
Mercurial barometers, 69–70
Mesa Verde, Colorado, 375
Mesoclimates, 5
Mesopause, 9
Mesophytic forests, 228
Mesopotamia, 254
Mesosphere, 8–9
Metabolic heat, 348
Meteorological tables, abridged, 411–15
Meteorology, defined, 4
Meteors, 9
Methane, 7
Meuse Valley, Belgium, 360
Mexico, 169, 375
Miami, Florida, 180
Microbarographs, 71
Microclimates, 5, 144, 212
  in forests, 291
  and housing, 331–34
  modification of, 384–85
  and soils, 235, 238
Mid-latitude arid climate, 188–92
Mid-latitude desert climate, 190
Mid-latitude semiarid climate, 188–92
Mid-latitude steppe climate, 190
Millibar, 69
Minnesota, 191
Mississippi Valley, 195
Missouri, 125, 198
Mist, 58
Mistral, 87, 176
Mixing fog, 56
Mixing ratio, 36
Mobile, Alabama, 180
Modification of weather and climate, 384–404
Modified humid continental climate, 199
Modifying microclimates, 384–85
Moisture, 35, 39, 150–51 (*See also* Precipitation; Rainfall)
  adequacy, 150–51
  atmospheric, 247–49
  crops and, 266–67, 280–86
  and electric transmission, 318
  and plants, 222–23
  soil and, 222, 236–41, 249–53, 280–83
Moisture index, 150
Monsoon region, 96
Monsoons, 64, 96–97, 161–63, 167, 180, 182
  drought and, 283
Monsoon tropics, 160–64
Montgomery, Alabama, 193
Moscow, U.S.S.R., 152, 197–98
Mountain barriers, 63–64, 182, 185, 212–13
  air masses and, 127–29
  in arid climates, 170
  fronts and, 115–16
  highland climates and, 212–13
  marine climates and, 185
  mid-latitude climates and, 188
  precipitation and, 63–64, 170, 182, 185
  and temperature, 128–29, 185
Mountain breezes, 87
Mountain wave, 308
Mt. Rainier National Park, Washington, 64
Mt. Tamalpais, California, 80–81
Mt. Waialeale, Hawaii, 59
Mt. Washington, New Hampshire, 80
Muskingum River, 259

N

Nagpur, India, 166
Namib Desert, 169
Natal Coast, 178
National Academy of Sciences Panel on Modification of Weather and Climate, 400
National Oceanic and Atmospheric Administration, 401
National Weather Service, 305
Natural gas, 319
Navigation and weather, 311–15
Nebraska, 194, 275, 277–78, 280
Nemuro, Japan, 199–200
Neon, 7
Nephanalysis, 131
Nephoscope, 52

Irish Sea, 188
Irkutsk, Siberia, 149
Irrigation, 284–86
Isobaric slope, 76
Isobars, 72
Isohyets, 63
Isophenes, 275–76
Isotachs, 84
Isothermal lapse rate, 33
Isotherms, 27
Ivigtut, Greenland, 208

J

Jacobabad, Pakistan, 170
Japan, 118, 127, 178, 180, 198–200, 228, 268
Jet streams, 92–93, 97, 162
  air routes and, 306
  turbulence and, 309
Jidda, Arabia, 282–83
"Jungle," 227

K

Kamchatka Peninsula, 202
Kansas, 280
Kashmir, 214
Katabatic winds, 87, 176, 212
Katmai, 380
Kauai, 59
Kelvin scale, 23
Kendall, Henry M., 191
Kendrew, W. G., 152
Kenya, 170
Kernlose winter, 210
Khamsin, 87
Khasi Hills, 59, 182
Klima, 4
Klimata, 142
Knots, 81–82
Koeppen, Wladimir, 145
Koeppen classification, 145–48, 154, 185
Korea, 193
Krakatoa, 93, 380
Krakatoa easterlies, 93
Krypton, 7
Kuala Lumpur, Malaya, 158

L

Labrador Coast, 209
Labrador Current, 200
Lahore, Pakistan, 171
Land breeze, 87
La Paz, Bolivia, 215
Lapse rates, 32, 104–6
Latent heat of condensation, 42, 44, 127
Latent heat of fusion, 43
Latent heat of vaporization, 43
Latin America, 233 (*See* also country names)
Latitude:
  climate and, 156, 164, 168, 173, 182, 188, 192–93, 196, 202–3, 206–7, 210, 212, 214
  climatic classification and, 142
  and duration of daylight, 16–18
  frost-free season and, 274
  insolation and, 17–18
  jet streams and, 91–92
  and seasons, 17–18
  and snowline, 216
  and transparency of atmosphere, 16–17
  wind direction and, 77–78, 88–89
Leaching, 236
Leh, Kashmir, 214
Leveche, 87
Libby, Willard F., 371
Libya, 170
Lidar, 53
Lightning, 124
  and cloud modification, 400
  and forest fires, 292
Lightning protection, 340
Lightning rods, 340
Lincoln, Nebraska, 276–77
Lithosphere, 3
"Little Ice Age," 375
Little Rock, Arkansas, 180
Livestock, and climate, 293–300
Llanos, 229
Local relief, 212, 241, 331
Local winds, 86–88
  and air pollution, 324
  and house sites, 331
Locust swarms, 298–99
Lodging, of grains, 280, 286
London, England, 360
Longs Peak, Colorado, 214
Long-wave radiation, 12–15, 381
Louisiana, 180, 193
Lovelock, Nevada, 189
Luxor, Egypt, 171

M

Mackenzie River, 256
McMurdo Sound, Antarctica, 210
Macroclimates, 5, 144
  and housing, 331
  modification of, 389–92
Madagascar, 156, 159, 164
Madras, India, 165
Maine, 199
Malabar Coast, 162
Malakal, Sudan, 168
Malnutrition, 357

Nevada, 189
New England, 197, 199–200
"New England-Hokkaido" climate, 199
Newfoundland, 199–200
New Guinea, 156, 212
New Hampshire, 80
New Orleans, Louisiana, 179–80, 193
New Zealand, 63, 88, 182, 184–85, 300
Nile flood plain, 254
Nile floods, 375
Nimbostratus, 48
NIMBUS III photograph, 119
Nitric oxide emissions, 392
Nitrogen, 7
Nitrogen oxides, 359
Nitrous oxide, 7
Nonadiabatic processes, 44
Normal lapse rate, 32
North Africa, 63, 99, 169, 298 (*See also* country names)
North America, 182, 185, 194, 196, 202, 206, 212, 365 (*See also* Canada; United States)
North Atlantic drift, 31, 183, 208
North Atlantic Ocean, 73, 188, 208–9
North Carolina, 402
North Dakota, 190–91
Northeast trades, 88
Norther, 88
North Sea, 375
Norway, 202–3, 207
Nor'wester, 88
Nuclear explosions, 392
Nucleation, artificial, 398

O

Oakland, California, 179
Ob River, 256
Ocean currents, 63, 97, 169
  and climatic change, 381
  and humid subtropical climate, 178
  map of, 98
  and modification of climate, 396
Ocean spray salts, 42
Occluded fronts, 113–16
Occlusions, 113–16
Ohio, 259
Ohio Valley, 255
Oimekon, Siberia, 204
Oklahoma, 280
Olympic Mountains, 216
Omaha, Nebraska, 194
Ordovician period, 369
Oregon, 183
Orinoco Valley, 229
Orographic lifting, 43–44
Orographic precipitation, 59, 159
Otago Basin, 185
Overhangs, 335–36, 338
Oxygen, 7
Ozone, 7–8, 359, 381, 392

P

Pacific Coast, 183, 185
Pacific high, 129, 183
Pacific Northwest, 376
Pakistan, 169, 171
Paleoclimates, 369–70
Paleoclimatology, 369–70
Paleozoic glaciation, 369
Palouse region, 242
Pampas, 230
Pampero, 88
Panel heating, 342
Paradise Ranger Station, Washington, 64
Paraguay, 164
Parasites, 299
Paris, France, 184
Patagonia, 189, 192, 229
Peat bogs, 374
Pedogenic criteria, 374
Penman, H. L., 252
Perception of climate, 365
Perihelion, 16
Permafrost, 231–32
  and floods, 258
Perspiration, 348, 354–55
Peru, 169, 172
Peru Current, 31, 98, 169
Peru Desert, 169
Phenological data, 275–79
Phenological gradient, 275
Phenology, 275–79
Philippines, 156, 161
Photochemical reactions, 323
Photoperiod, 279 (*See also* Daylight)
Photosynthesis, 223, 277
Physical climatology, 4
"Pibal," 82
Pigs, 295, 297
Pilot balloons, 82
Plant growth:
  climate and, 221–25
  heat and, 222–25
  humidity and, 222
  moisture and, 222–23, 280
  sunshine and, 223–24
  temperature and, 222–24, 266–71
  wind and, 224, 286–87
Plant introduction, 300–301
Plass, Gilbert N., 389
Pleistocene epoch, 370
Poland, 197

Polar air masses, 108, 153, 175, 193, 197, 203
Polar climate, 209–12
Polar easterlies, 90, 109
Polar front jet stream, 92
Polar fronts, 90
Polar highs, 73, 90
Polar marine climate, 206
Polar night jet stream, 93
Pollen studies, 374
Pollution, air, 322–26, 331, 359–60
Portland, Oregon, 183
Portugal, 182
Potential evapotranspiration, 148–52, 165, 190, 206, 251–52
  measurements, 251–52
  radiation and, 252
Poultry, 295
Prairie, 228–30
Precession of the equinoxes, 380
Precipitation (*See also* Rainfall; Snow):
  air pollution and, 324
  in arid and semiarid climates, 171–72, 190–92
  in cities, 388–89
  construction activities and, 327–28
  defined, 56
  dependability, 64
  in dry summer subtropics, 176–77
  and floods, 254–57
  in forests, 289–90
  forms of, 58–59
  frequency of, 65
  in highland climates, 215–16
  house sites and, 332–33
  in humid continental cool summer climate, 198–200
  in humid continental warm summer climate, 194–95
  in humid subtropics, 180–82
  intensity of, 65–66
  livestock and, 294–96
  in marine climate, 185–87
  in mid-latitude arid and semiarid climates, 190–92
  in monsoon tropics, 160
  number of days, in United States, 187
  observations, 60–61
  in polar climate, 211
  in rainy tropics, 157–60
  regime, 65
  regional distribution of, 61–64
  reliability, in marine climate, 186; in wet-and-dry tropics, 166
  in taiga climate, 204, 206
  in tropical arid and semiarid climates, 171–72
  in tundra climate, 208–9
  types, 59–60
  variability, and drought, 281, 283
  in wet-and-dry tropics, 165–67
  world distribution, 61–64
Precipitation effectiveness, 148
Pressure gradient, 76–78
Pressure-height relationships, 71–72
Pressure pattern flying, 306
Pressure scales, 69–70
Prevailing westerlies, 89–90
Profile, of soil, 237, 240
Project Cirrus, 401
Project Skyfire, 401
Project Stormfury, 401
Pseudoadiabatic rate, 44
Psychrometers, 37–38
Public utilities, 318–21
Puerto Rico, 161
Puget Sound-Willamette Lowland, 185
Punta Arenas, Chile, 184–86
Pyranometers, 19–20
Pyrheliometers, 20

Q

Quito, Ecuador, 214

R

Radar:
  and snow surveying, 261
  wind observations by, 84
Radar storm detection, 130
Radiation, 13–15, 252
  budget, 12–15, 20
  definition, 13
  infrared, 12
  long-wave, 7, 12, 14–15, 381
  short-wave, 12–14, 381
  ultraviolet, 12, 357, 379, 381
  in wet-and-dry tropics, 165
Radiation fog, 54–55, 304
Radiation frosts, 269–73
Radio, and ionosphere, 10
Radio communications, 321
Radioisotope dating, 370–71
Radiometers, 19–20
Radiosondes, 101–2
Radon, 7
Railways, weather and, 315–16
Rain, 56–58 (*See also* Rainfall)
Rainfall (*See also* Precipitation):
  dependability, 64–65
  frequency, 64–65
  intensity, 65–66
  seasonal variation of, 64–65
Rain gauges, 60–61
Rainmaking, 249, 384, 400

Rainy tropics, 156–60
Raobs, 101
Rawin observations, 84
Rawinsonde, 101
Red Sea, 394
Regime of precipitation, 65
Regional climatology, 4–5, 140
Relative humidity, 36–39
  air conditioning and, 340–42, 345
  in cities, 389
  comfort and, 342, 348–50
  cooling power and, 350
  defined, 37
  disease and, 363
  effective temperature and, 341–42, 349–50
  in forests, 290
  health and, 363
  latitude and, 36–37
  sensible temperature and, 349
Rhone Valley, 87
Rime, 59, 310
Riviera, 87
"Roaring forties," 89, 188
Robinson cup anemometer, 82
Rockets, 131, 393, 401
Rocky Mountains, 212, 216
Romania, 193–94
Rossby waves, 91
Ross Sea, 210
Runoff, 254–58
  in Canada, 256–57
  in Florida, 256
  from houses, 338
  in United States, 256–57
  in urban areas, 389
Runoff forecasts, 258
  snow surveying and, 260–62
  temperature and, 259

S

Sacramento, California, 175–76
Sahara, 87, 169, 229, 394
Sailplanes, 306–7
St. Louis, Missouri, 193, 198
Sakhalin, 197
Salvador, Brazil, 160
San Francisco, California, 80, 152, 175–76
San Juan, Argentina, 190–91
Santa Ana, 176
Santa Cruz, Argentina, 189, 191
Santiago, Chile, 176
Santis, Switzerland, 215
Sastrugi, 211–12
Satellites, 131
Saturation, 36
Saturation vapor pressure, 36
Savanna, 228–29
Savannah, Georgia, 402
Scandinavia, 197, 373, 377
Scandinavian Peninsula, 31, 182
Schaefer, Vincent J., 397
Sclerophyllous forest, 227
"Scotch mist," 58
Scotland, 327
Scrub forest, 226–27
Sea breeze, 86–87
Sea ice, 206, 375
Sea level, 370–71
Sea level pressure, 69, 72–75
"Sea smoke," 54
Sea swell, 121, 314
Selva, 226
Semiarid climates, 168–72, 188–92
Semideciduous forest, 227
Sensible temperature, 349
Sewage systems, 321
Shear line, 163
Sheep, 295, 297
Shillong, India, 182
Short-wave radiation, 11–14, 17, 381
Siberia, 189, 197, 203, 269, 396
Sierra Nevada, 212, 234–35
Silurian, 369
Silver iodide, 43, 398–99
"Silver thaw," 58
Simpson, Sir George, 378–79
Singapore, 157, 171
Sirocco, 87
Sitka, Alaska, 186
Sky cover, 51–52
Sleet, 58
Sling psychometer, 37–38
Smog, 323, 359
Smudge pots, 271
Snag, Yukon Territory, 204
Snow, 56–59 (*See also* Precipitation)
  avalanches of, 262–63
  depth of, 61
  in dry summer subtropics, 177
  and floods, 255–56
  in forests, 289–90
  in highland climates, 214–16
  highways and, 316–18
  house construction and, 339
  in humid continental cool summer climate, 198
  in humid continental warm summer climate, 196
  in humid subtropics, 182
  in marine climate, 187
  in polar climate, 209–11
  in taiga climate, 204, 206
  in tundra climate, 209
Snow courses, 260–61

Snow fences, 318
Snowflakes, 58
Snowline, 215-16
Snow load, 339
Snow pellets, 59
Snow pillow, 261
Snow removal, 317-18
Snow sampler, 260–61
Snow sheds, on highways, 318
Snow surveying, 260–62
Soil Conservation Service, 239, 260
Soil erosion, 242–44
Soil horizons, 237–40
Soil moisture, 222, 236–41, 249–53, 280–83
  field capacity, 252–53
  and wilting point, 252–53
Soil moisture tensiometer, 253
Soil profile, 237–40
Soils:
  classification, 238–39
  climate and formation of, 235–42
  of highland regions, 241–42
  relief and, 235, 238, 241
  spatial distribution of, 239–42
Solar constant, 16, 378
Solar heating, 336
Solar radiation, 12, 18, 323, 336, 378 (*See also* Insolation)
Solar spectrum, 12, 16, 323
Solar variations, 378–80
Somalia, 169
Sonoran Desert, 169
Source regions, of air masses, 100–101, 106–8
South Africa, 178, 228–30
South America, 156, 161, 164, 229, 300 (*See also* country names)
South Atlantic, 118
South Australia, 175, 177
Southeast Asia, 64, 163–64, 226, 300
Southeast trades, 88
Southern Alps, 63–64
South Pole, 210
Southwest Africa, 169
Southwest Asia, 298
Spain, 87
Special observations, 305
Specific heat, 30
  of water, 246
Specific humidity, 35–36
Springfield, Colorado, 191
Squall line, 123
Stable air, 104–5, 116
  and air pollution, 323–25, 359–60
Stable-air foehn effect, 128–29, 214
Station pressure, 69
Steam fog, 54
Steinbeck, John, quoted, 365–66
Steppe, 228–29
Steppe climate, 190
Steppe vegetation, 190
Storms:
  electrical, 130
  flying in, 310–11
  modification of, 401-2
  observations of, 130–32
  in rainy tropics, 158
  tropical, 117-21
Storm surges, 121, 259
Strait of Juan de Fuca, 185
Straits of Magellan, 186, 188
Stratocumulus, 49
Stratopause, 9
Stratosphere, 8–9
  dust in, 392
  volcanic ash in, 380
Stratus, 49
Stream lines, 84
Subarctic climate, 202, 374
Sublimation, 43
Sublimation nuclei, 43
Subpolar low, 73, 90
Subsidence, 76–77, 104, 168, 188
Subsidence inversions, 33, 168, 215
Subtropical jet stream, 92
Subtropic highs, 73, 88–89
Sudan, 229
Sulphur dioxide, 359–60
Sumbawa, 380
Sunda, Strait of, 380
Sunlight, in forests, 289
Sun scald, 269
Sunshine:
  and health, 357–59
  and house orientation, 333–34
  measurement, 18–20
  and plant growth, 223
  and tourist resorts, 359
Sunspots, 378–79
Supercooled clouds, 309, 397–400
Supersaturation, 42, 397
"Surge of the trades," 163
Suva, Fiji, 160
Swiss Avalanche Commission, 262
Switzerland, 18, 215
Sylphon cell, 70–71
Synoptic observations, 133
Synoptic weather chart, 133

T

Taal, 380
"Table Cloth," 48
Tacoma, Washington, 327
Tacoma Narrows Bridge, 327
Taiga, 202, 228

Taiga climate, 202–6
Tailwinds, 306
Taiwan, 178
Tambora, 380
Tasman Sea, 188
Tay Bridge, 327
Telegraph services, 320
Telephone services, 320
Temperature, 4, 8–10, 21–32
  air conditioning and, 296, 340–46
  altitude and, 9–10, 32–33, 214
  animals and, 294–97
  in cities, 385–89
  crops and, 266–71
  cyclical variations, 377–78
  daily mean, 25
  diurnal range, 25
  in dry summer subtropics, 175–76
  effective, 341–43, 349–50
  equivalent wind chill, 351
  extremes, table, 27
  forest-fire danger and, 292
  in forests, 292
  globe, 342
  health and, 355–57
  in highland climates, 214–15
  horizontal distribution of, 27–31
  house design and, 335–37
  human comfort and, 341–42, 347
  in humid continental cool summer climate, 197–98
  in humid continental warm summer climate, 193–94
  in humid subtropics, 178–80
  indoor climate and, 340–46
  insects and diseases and, 298–300
  lapse rates, 32–33
  livestock and, 294–97
  lowest observed in Northern Hemisphere, 205
  in marine climate, 183–84
  mean radiant, 342
  measurement, 21–25
  in mid-latitude arid and semiarid climates, 188–90
  in monsoon tropics, 161
  plant growth and, 222–24, 266–71
  in polar climate, 209–11
  in rainy tropics, 156–57
  records, 25–27
  response of body to, 348–52, 355–57, 363–64
  runoff forecasting and, 259
  sensible, 349
  in taiga climate, 203–4
  trends, in United States, 376–77, 382
  in tropical arid and semiarid climates, 168, 170–71
  in tundra climate, 207–8
  vertical distribution, 32–33
  in wet-and-dry tropics, 165
  world trends, 377
Temperature efficiency, 148, 150
Temperature-humidity index, 350
Temperature inversions, 32–33
  air pollution and, 323–25, 360
  house sites and, 332
Tennessee, 180
Texas, 88, 280
Thames Valley, 360
Thawing, 223, 243–44 (*See also* Melting)
Theodolite, 82, 84
The Pas, Manitoba, 283
Thermal belt, 270–71
Thermal convection, 9, 78
Thermal efficiency, 149–50
Thermistors, 23, 82, 101
Thermocouples, 23, 82
Thermographs, 24–25
Thermometers, 23–25
  dry-bulb, 37
  maximum, 23–24
  minimum, 24
  wet-bulb, 37–39
Thermosphere, 8–10
Thom, E. C., 349
Thorn forest, 226
Thornthwaite, C. W., 148–50, 252
Thornthwaite classifications, 148–52
Thunder, 124
Thunderhead clouds, 51, 123
Thunderstorms, 121–25
  air mass, 124
  annual number in world, 121
  floods and, 255–56
  forest fires and, 292
  frontal, 124
  in humid continental cool summer climate, 198
  in humid continental warm summer climate, 194, 196
  in humid subtropics, 180, 182
  in mid-latitude arid and semiarid climates, 192
  observations of, 124
  in rainy tropics, 158
  soil erosion and, 243
  squall line in, 123
  in tropical deserts, 172
  in United States, 181
  in wet-and-dry tropics, 165
Tibet, 212
Tierra caliente, 233–34
Tierra fria, 234
Tierra helada, 234
Tierra templada, 234

Tipping-bucket rain gauge, 60–61
TIROS IV photograph, 50
Tokyo, Japan, 180–81, 282–83
Topoclimate, 144
Topoclimatology, 5
Tornadoes, 124–27, 192, 195
  damage by, 125
  frequency, 126–27
  paths of, 125–27
Tornado proofing, 338
Trace, of precipitation, 60
Trade-wind inversion, 168, 215
Trade winds, 88, 156, 163, 170
Transmissometer, 54–55
Transosonde, 101–2
Transpiration, 40
Transportation:
  highway, 316–18
  railway, 315–16
  water, 311–15
Tree-rings, 372–73, 375
Trewartha, Glenn T., 145
Trivandrum, India, 161
Tromso, Norway, 203
Tropical air masses, 107–8, 156, 174, 178, 193, 197
Tropical arid climate, 168–72
Tropical crops, 300
Tropical rain forest, 226
Tropical semiarid climate, 168–72
Tropical storms, 117–21
Tropopause, 9
Tropophytes, 222
Troposphere, 8, 11
Troughs, of low pressure, 116–17
Trujillo, Peru, 172
Tundra, 206, 231–33
Tundra climate, 206–9
Tundra soils, 239
Tundra vegetation, 231–33
Turbulence, 56, 78–79
  aviation and, 308–9
  fog and, 55
Twilight, 8, 208
Typhoons, 118, 180
Typhoon season, 200

U

Ultraviolet radiation, 14, 357, 379, 381, 392
United States:
  broadleaf-coniferous forest in, 228
  chinooks in, 127
  climatic trends in, 373, 376–77, 382
  dendroclimatology in, 373
  "Dust Bowl" in, 244
  flood forecasting in, 259
  floods in, 256
  forest fires in, 292, 401
  Gulf Stream and, 63
  hail in, 280–81
  humid continental warm summer climate in, 192–93
  humid subtropical climate in, 178
  "Indian summer" in, 196
  maximum temperature in, table, 27
  mid-latitude arid and semiarid climates in, 188
  precipitation in, 196
  precipitation gauge stations in, 60
  rain forest in, 228
  temperatures in, 31, 193–94
  thunderstorms in, 181
  tornadoes in, 125–26
  tropical arid climate in, 169
  waterspouts in, 127
  weather maps of, 134, 136
  weather stations in, 315, 376
Updrafts, 308
Upper-air charts, 134
Upslope fog, 56
Upwelling, 99, 169
  and modification of climate, 396
Urban aerosol, 385–88
Urey, Harold C., 370
U.S.S.R., 188, 197, 200, 203–4, 228, 230

V

Vadose water, 253
Valley breezes, 87
Valparaiso, Chile, 175–76
Vaporization, heat of, 40, 43
Vapor pressure, 36
  and plants, 222
Vardo, Norway, 207–8
Varves, 371–72
Vegetation, 221–35 (*See also* Plant growth)
  climax, 226
  of deserts, 230–31
  of grasslands, 228–30
  in mid-latitude arid and semiarid climates, 190–92
  in tundra, 231–33
  vertical differentiation of, 233–35
  world distribution of, 225–35
Ventilating systems, 342, 344
Venting, 327, 336, 338
Vents, in structures, 327
Verkhoyansk, Siberia, 204
Vertical life zones, 233–35
Vertical zonation, of soils, 241–42
Vicksburg, Mississippi, 180
Victoria, British Columbia, 184
Vietnam, 161
Vikings, 375

Virga, 48
Visibility, 54
  air pollution and, 322
  aviation and, 304–5, 310
  and water transportation, 313
Vladivostok, U.S.S.R., 199–200
Volcanoes, 380
Vostok II, Antarctica, 210

W

Walvis Bay, Southwest Africa, 169
Warm fronts, 111–12
Warm sector, 110
Washington, D.C., 65
Washington State, 187, 327
Water:
  changes in state of, 39–43
  in ice caps, 368
Water balance equation, 40
Water budget, of the atmosphere, 35
Water budget graphs, 149, 282
Water cycle (*See* Hydrologic cycle)
Water droplets, 8
  in clouds, 45
  and condensation, 42
  and frost prevention, 273
  and precipitation, 57–58, 397–400
Water need, 148, 249, 252, 283–85 (*See also* Potential evapotranspiration)
Waterspouts, 124–26
Water vapor, 7, 35–43, 249
  and climatic change, 381
  in thermosphere, 392
Wave-induced turbulence, 309
Weather:
  construction and, 326–28
  definition, 3
  health and, 355–57
  industry and, 321–28
  labor force and, 322
  manufacturing and, 321–22
  public utilities and, 318–21
Weather codes, 133
Weather forecasting, 133–37
  computers and, 133, 137
Weather house, 39
Weather maps, 132–36
Weather modification, 384–404
  implications of, 404
Weather observations, 137
  synoptic, 133
Weather observing stations, 133
Weather planes, 130–31
Weather records, 375–76
Weather ships, 144
Weather stations, mobile, 293
Weighing-type rain gauge, 60–61
Westerlies, 89–91, 111, 117, 175, 183, 188
Westerly jet streams, 162, 306
Wet adiabatic rate of cooling, 44
Wet-bulb depression, 37
Wet-bulb temperature, 37, 349–50
Wet-and-dry tropics, 164–68
Whirlwinds, 129
"Whiteout," 204
Willett, Hurd C., 379
Willy-willies, 118
Wilting point, 252–53
Windbreaks, 287–88, 297, 395
Wind chill, 351–52
Wind direction frequency polygon, 85–86
Wind erosion, 243–44
Windfall, 291
Windhoek, Southwest Africa, 169
Wind register, 81
Wind roses, 85–86, 175, 200, 207
Winds:
  air pollution and, 323–25, 360, 362
  allergens and, 360
  in cities, 388–89
  crops and, 286–88
  direction of, 81–82
  in dry summer subtropics, 176
  factors affecting direction and speed of, 77–78
  in forests, 290–91
  friction and, 78, 80
  in highland climates, 213
  house orientation and, 333–34
  house sites and, 331–33
  in humid continental cool summer climate, 199–200
  insect migration and, 298–99
  livestock and, 294, 296
  local, 86–88
  in marine climate, 188
  mean flow patterns, 94–96
  in mid-latitude arid and semiarid climates, 192
  in monsoon tropics, 162–64
  observations, 80–84
  in polar climate, 211–12
  pressure gradient and, 76–79
  in rainy tropics, 159–60
  structures and, 327
  in tropical arid and semiarid climates, 168–69
  vegetation and, 224
Winds aloft, 79–80
Wind shear, 79, 116, 309
Wind shift, 113, 116
Wind speed:
  altitudinal variation, 79
  diurnal variation, 80
  friction and, 78–80

  in jet streams, 92–93
  measurement, 80–84
Wind vanes, 81–83
Wind velocity, 82 (*See also* Wind speed)
Winnipeg, Manitoba, 197–98
WIT, 309
Woodland, Maine, 199
World Meteorological Organization, 45, 133
World Meteorological Organization Code, 133
Wyoming, 280

X

Xenon, 7
Xerophytes, 222, 230

Y

Yakutsk, Siberia, 203
Yellowknife, N.W.T., 126
Yemen, 170
Yenesei River, 256
Yugoslavia, 87, 176, 193
Yukon Territory, 204

Z

Zero temperatures, 277
Zonal circulation, 379–80
Zonal soils, 238
Zonda, 192
Zones of convergence, 101
Zurich, Switzerland, 282–83